

ENSO Asymmetry in CMIP5 Models

TAO ZHANG AND DE-ZHENG SUN

Cooperative Institute for Research in Environmental Sciences, University of Colorado, and NOAA/Earth System Research Laboratory/Physical Sciences Division, Boulder, Colorado

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ABSTRACT

The El Niño–La Niña asymmetry is evaluated in 14 coupled models from phase 5 of the Coupled Model Intercomparison Project (CMIP5). The results show that an underestimate of ENSO asymmetry, a common problem noted in CMIP3 models, remains a common problem in CMIP5 coupled models. The weaker ENSO asymmetry in the models primarily results from a weaker SST warm anomaly over the eastern Pacific and a westward shift of the center of the anomaly. In contrast, SST anomalies for the La Niña phase are close to observations.

Corresponding Atmospheric Model Intercomparison Project (AMIP) runs are analyzed to understand the causes of the underestimate of ENSO asymmetry in coupled models. The analysis reveals that during the warm phase, precipitation anomalies are weaker over the eastern Pacific, and westerly wind anomalies are confined more to the west in most models. The time-mean zonal winds are stronger over the equatorial central and eastern Pacific for most models. Wind-forced ocean GCM experiments suggest that the stronger time-mean zonal winds and weaker asymmetry in the interannual anomalies of the zonal winds in AMIP models can both be a contributing factor to a weaker ENSO asymmetry in the corresponding coupled models, but the former appears to be a more fundamental factor, possibly through its impact on the mean state. The study suggests that the underestimate of ENSO asymmetry in the CMIP5 coupled models is at least in part of atmospheric origin.

1. Introduction

The El Niño–Southern Oscillation (ENSO)—a major source for interannual climate variability—affects weather and climate worldwide (Ropelewski and Halpert 1987; Kiladis and Diaz 1989; Hoerling et al. 1997; Larkin and Harrison 2005; Sun and Bryan 2010; Zhang et al. 2011, 2014). The two phases of ENSO—El Niño and La Niña—which are defined as tropical Pacific anomalies relative to a long-term average, are not mirror images of each other: the strongest El Niño is stronger than the strongest La Niña, a fact that has been referred as ENSO asymmetry (Burgers and Stephenson 1999).

The asymmetry between two phases of ENSO shows up in both the surface fields as well as in the subsurface fields (Rodgers et al. 2004; Schopf and Burgman 2006; Sun and Zhang 2006; Zhang et al. 2009). Causes for such an asymmetry are not yet clearly understood, but many studies suggest that it is likely a consequence of

nonlinearity of the ocean dynamics (Jin et al. 2003; An and Jin 2004; Su et al. 2010). By the analysis of the heat budget of the ocean surface layer, Jin et al. (2003) and An and Jin (2004) found that the nonlinear vertical temperature advections are a major contributor to the ENSO amplitude asymmetry. However, based on the updated ocean assimilation products, Su et al. (2010) suggested that the nonlinear zonal and meridional ocean temperature advections are essential to cause the asymmetry in the far eastern Pacific, while the vertical nonlinear advection has the opposite effect. Another possible cause for the ENSO asymmetry is the asymmetric negative feedback due to the tropical ocean instability waves in the eastern Pacific that has a relatively stronger impact on the La Niña than El Niño (Vialard et al. 2001). Kang and Kug (2002) argued that the relatively weak SST anomalies during La Niña compared to those during El Niño result from the westward shift of zonal wind stress anomalies during La Niña relative to El Niño. Such an asymmetry in the zonal wind stress between two phases of ENSO is in turn attributed to the nonlinear dependence of deep convection on the SST (Hoerling et al. 1997). A recent review paper by

Corresponding author address: Dr. Tao Zhang, NOAA/ESRL/PSD, 325 Broadway, R/PSD1, Boulder, CO 80305.
E-mail: tao.zhang@noaa.gov

TABLE 1. List of the 14 models used in this study and their expanded names.

Model	Expanded name
BCC-CSM1-1	Beijing Climate Center, Climate System Model, 1-1
CCSM4	Community Climate System Model, version 4
CNRM-CM5	Centre National de Recherches Météorologiques Coupled Global Climate Model, version 5
CSIRO Mk3.6.0	Commonwealth Scientific and Industrial Research Organisation Mark, version 3.6.0
FGOALS-g2	Flexible Global Ocean–Atmosphere–Land System Model gridpoint, version 2
FGOALS-s2	Flexible Global Ocean–Atmosphere–Land System Model gridpoint, second spectral version
GISS-E2-R	Goddard Institute for Space Studies Model E, coupled with Russell ocean model
HadGEM2-ES	Hadley Centre Global Environmental Model 2, Earth System
INM-CM4	Institute of Numerical Mathematics Coupled Model, version 4.0
IPSL-CM5A-LR	L’Institut Pierre-Simon Laplace Coupled Model, version 5, coupled with NEMO, low resolution
MIROC5	Model for Interdisciplinary Research on Climate, version 5
MPI-ESM-LR	Max Planck Institute Earth System Model, low resolution
MRI-CGCM3	Meteorological Research Institute Coupled General Circulation Model, version 3
NorESM1-M	Norwegian Earth System Model, intermediate resolution

An (2009) provides a good account of the aforementioned theories for ENSO asymmetry. In a more recent study by Liang et al. (2012) using an analytical model of Sun (1997), it is noted that ENSO asymmetry may depend on the radiative forcing as in that model a stronger radiative forcing produces a stronger and more positively skewed oscillation. They also attribute the asymmetry of the two phases of ENSO—as traditionally defined as the deviations from the climatological mean—to the asymmetry of the dynamics relative to the equilibrium state of the system.

Understanding the causes and consequences of ENSO asymmetry may hold the key to understand decadal variability in the tropics and beyond, as the asymmetry suggests a time-mean effect of ENSO (Rodgers et al. 2004; Schopf and Burgman 2006). Indeed, in theoretical studies and numerical experiments designed to determine the time-mean effect of ENSO, an association between the time-mean effect of ENSO and the asymmetry of ENSO is found (Sun and Zhang 2006; Sun and Yu 2009; Sun et al. 2013; Sun et al. 2014), although it appears that they are both a consequence of the nonlinearity. To fully capture the role of ENSO in the climate system, the climate models need to simulate well the asymmetry of ENSO.

The ENSO asymmetry in coupled models has been extensively examined in previous studies (Burgers and Stephenson 1999; Hannachi et al. 2003; An et al. 2005; van Oldenborgh et al. 2005; Zhang et al. 2009; Sun et al. 2013). The studies of van Oldenborgh et al. (2005) and Sun et al. (2013) made use of the archive of the models from phase 3 of the Coupled Model Intercomparison Project (CMIP3) and found that an underestimate of the asymmetry is a prevalent problem, capping the early findings from a rather scattered set of models. However, the cause for the bias in ENSO asymmetry is not well

understood in those studies. In the complex coupled system it is difficult to identify causes for biases in ENSO asymmetry owing to the strong feedbacks of the ocean–atmosphere system in the tropical Pacific. Understanding the bias in coupled models therefore requires the use of component models, such as stand-alone atmospheric models, through which we can isolate the sources and amplifiers of biases in climate models.

In the present study, we evaluate the ENSO asymmetry in CMIP5 models (Taylor et al. 2012; see Table 1 for expanded model names of the models analyzed here). We follow the methodology of Zhang et al. (2009) and analyze the corresponding Atmospheric Model Intercomparison Project (AMIP) runs as well in order to gain more insight into the possible causes of the bias in ENSO asymmetry. By analyzing previous National Center for Atmospheric Research (NCAR) coupled models (CCSM1, CCSM2, and CCSM3 at T42; CCSM3 at T85; and CCSM3 + NR) in conjunction with the corresponding AMIP runs, Zhang et al. (2009) showed that all the models underestimate the observed ENSO asymmetry, but CCSM3 + NR with the Neale and Richter convection scheme (Neale et al. 2008) has significant improvements over the earlier versions with the Zhang and McFarlane convection scheme (Zhang and McFarlane 1995). Enhanced convection over the eastern Pacific during the warm phase of ENSO appears to be the cause for the improvement. Zhang et al. (2009) also noted a warmer SST climatology in CCSM3 + NR in contrast to other versions. We will explore whether the underestimate of ENSO asymmetry remains a common problem in the state-of-the-art coupled model; whether the underestimate of ENSO asymmetry in CMIP5 models is related to the weaker convection over the eastern Pacific during warm phase; and whether the mean SST state is important to ENSO asymmetry.

This paper is organized as follows: We introduce the observational and model datasets in [section 2](#). We present the analysis of ENSO asymmetry in CMIP5 models in the coupled runs, and then the asymmetry in the corresponding AMIP runs. To understand the impact of the biases identified from the analysis of the AMIP runs, the numerical experiments forced by AMIP winds are examined in [section 3](#). Conclusions and discussions are presented in [section 4](#).

2. Data and methods

The ENSO asymmetry in 14 coupled ocean–atmosphere models from CMIP5 control runs (piControl) has been evaluated in this investigation. Presented here are the results from the coupled models whose corresponding AMIP runs are available for the analysis. We will first assess the ENSO asymmetry in the SST and then look at the asymmetry in upper-ocean temperature in the models. We further analyze the corresponding fields of precipitation and surface wind stress in the coupled runs to understand whether the bias in ENSO asymmetry is linked to the bias in precipitation and associated surface wind stress. The corresponding AMIP runs from CMIP5 models are also examined to understand whether the biases in precipitation and surface wind stress in coupled runs stem from the biases in stand-alone atmosphere models.

In addition to analyzing the asymmetry in the CMIP5 AMIP runs, we also use the NCAR Pacific basin model to perform the forced ocean experiments driven by CMIP5 AMIP winds. Our model is the one used by [Sun \(2003\)](#), [Sun et al. \(2004\)](#), and [Sun and Zhang \(2006\)](#). The model uses the NCAR Pacific basin model ([Gent and Cane 1989](#)) as its ocean component. The model calculates the upper-ocean temperatures based on first principles and simulates well the observed characteristics of ENSO in both the forced and coupled modes ([Sun 2003](#)). We will compare the ENSO asymmetry in the runs forced by AMIP winds with that by observed winds to understand the effect of the bias in the atmospheric response on ENSO asymmetry in CMIP5 coupled models.

The observational data used for examining the model results are the same as those used by [Zhang et al. \(2009\)](#). The SST data from the Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST; [Rayner et al. 2003](#)) are used for evaluating the asymmetry in the SST field in the CMIP5 coupled models. The Simple Ocean Data Assimilation (SODA) dataset ([Carton et al. 2000](#)) is used for validating the upper-ocean temperature in the models. Precipitation data are obtained from the Climate Prediction Center (CPC) Merged Analysis

of Precipitation (CMAP; [Xie and Arkin 1997](#)). The wind stress data are obtained from the SODA dataset ([Carton and Giese 2008](#)) in which the surface winds are a combination of 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and Quick Scatterometer (QuikSCAT) satellite observations.

We will use the skewness ([Burgers and Stephenson 1999](#)) of interannual variability of SST to quantify the ENSO asymmetry. We will also conduct the composites of El Niño and La Niña and then use the sum of the composite between two phases of ENSO to measure the asymmetry. The definition of the warm phase and cold phase of ENSO follows that of [Zhang et al. \(2009\)](#). The composite analysis will help to identify which phase of ENSO the bias in ENSO asymmetry mainly originates from.

3. Results

a. Asymmetry in the coupled models

A quantitative measure of the ENSO asymmetry in CMIP5 coupled models reveals that an underestimate of the ENSO asymmetry remains a common bias in our state-of-the-art climate models. [Figure 1](#) shows the skewness of Niño-3 SST anomalies from observations and the models, together with their variance. Measured by the variance of Niño-3 SST, ENSO in many models is as strong as in observations. Measured by the skewness of Niño-3 SST, however, all the coupled models that we have analyzed underestimate the observed positive ENSO asymmetry. This indicates that the observed SST anomalies in the eastern Pacific are skewed toward warm events, while those in coupled models have a more Gaussian-like distribution. In comparison, the NCAR CCSM4 ([Gent et al. 2011](#); [Deser et al. 2012](#)) stands out as the best model in simulating the ENSO asymmetry, whose variability of ENSO is also comparable to observations. The HadGEM2-ES, which also has a comparable ENSO variability to observations, is found to have the largest bias in reproducing the observed positive skewness, because it shows a strong negative skewness, contrary to observations. The results suggest that the stronger variability of ENSO (measured by variance) does not guarantee a stronger asymmetry (measured by skewness) in CMIP5 coupled models.

[Figure 2](#) shows the sum of the SST anomalies between the warm and cold phases of ENSO from observations and coupled runs from CMIP5. This sum has also been called SST anomaly residual and is a common measure of the ENSO asymmetry in the SST field. The SST anomaly residual results are similar to the skewness map of SST anomalies (not shown). All the CMIP5 models

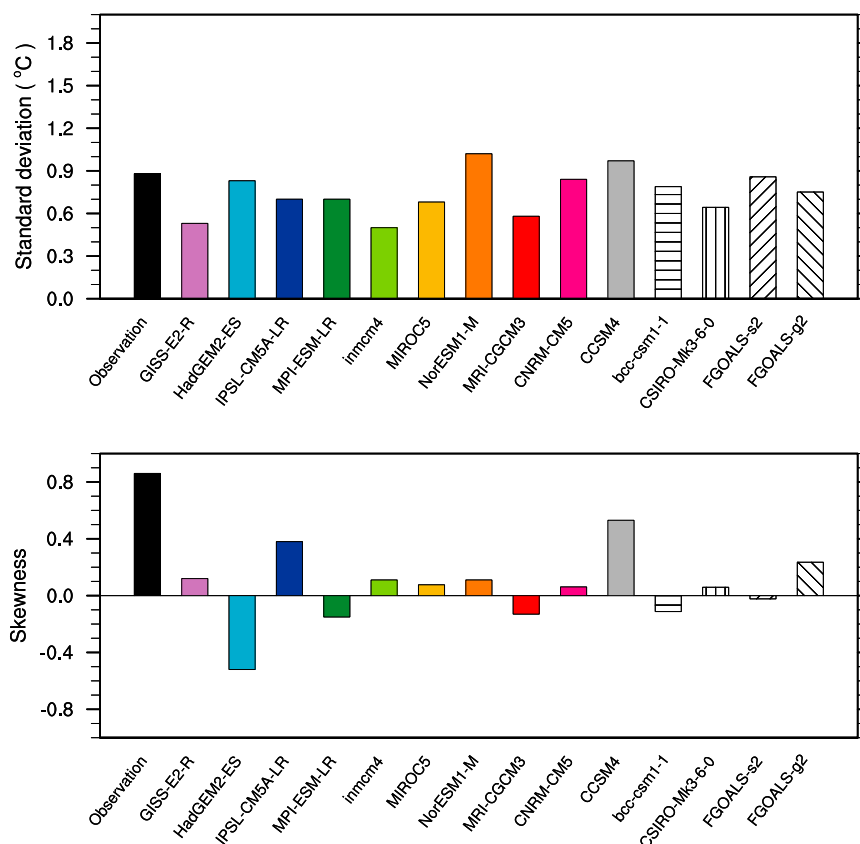


FIG. 1. (top) Standard deviation and (bottom) skewness of the interannual variability in Niño-3 SST from observations and CMIP5 coupled models. The length of data used in the calculation is 50 yr for all the models and observations (1950–99).

evaluated underestimate the observed positive SST residual and therefore the asymmetry over the eastern Pacific, consistent with the results of skewness. There is an obvious negative SST residual over the eastern Pacific in HadGEM2-ES, in agreement with a considerable negative skewness of Niño-3 SST anomalies in this model (Fig. 1). Generally, CCSM4 has a better simulation of the positive SST residual in the eastern Pacific than other models, which is also confirmed by the skewness results noted earlier. Despite the fact that all the models underestimate the positive SST residual over the eastern Pacific, the overestimate of the negative SST residual in the western Pacific is evident in many models (e.g., GISS-E2-R, MIROC5, CSIRO Mk3.6.0, CCSM4).

As already noted in the analysis of the previous NCAR models and consistent with earlier understanding of ENSO dynamics, the asymmetry in the subsurface temperature is more profound than in the surface (Zhang et al. 2009). To obtain more information about the cause for the bias in simulated ENSO asymmetry, we look at the asymmetry of the subsurface signal. Figure 3

shows the sum of the equatorial upper-ocean temperature anomalies between the warm and cold phases of ENSO from observations and coupled runs from CMIP5 models. The observed subsurface temperature shows a positive asymmetry of about 1°C around 75-m depth over the eastern Pacific and a negative asymmetry of about -0.4°C around 150-m depth over the western Pacific. All the models underestimate the positive asymmetry in the subsurface temperature over the eastern Pacific. In contrast to the asymmetry in SST, the underestimate of the positive asymmetry in the subsurface temperature is more profound over the eastern Pacific (note the different scales in Figs. 2, 3). Most models also have a weaker negative asymmetry in the subsurface over the western Pacific. Despite the comparable magnitude to observations, the negative asymmetry over the western Pacific extends too far to the east in some models (CNRM-CM5, FGOALS-g2, and CCSM4). There is a good match between SST and subsurface temperature for the negative asymmetry in HadGEM2-ES over the eastern Pacific (Figs. 3, 2). Consistent with the stronger positive SST residual over the eastern Pacific, CCSM4

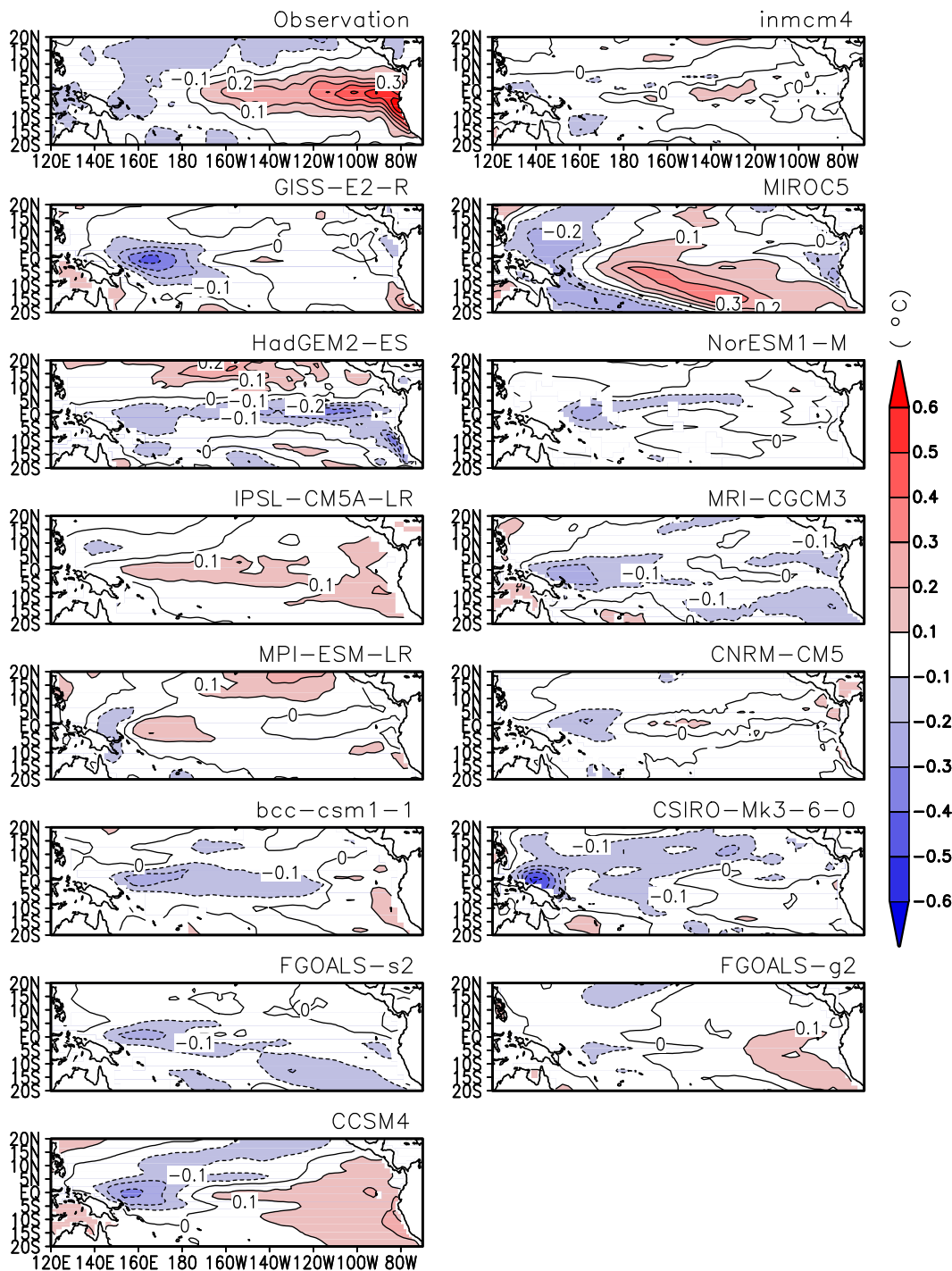


FIG. 2. The sum of the composite SST anomalies between the two phases of ENSO from observations and CMIP5 coupled models. Following the study of Zhang et al. (2009), the positive (negative) anomalies of Niño-3 SST with a value greater than 0.5°C (-0.5°C) are selected to construct composites of warm (cold) events. The same data are used as in Fig. 1.

also has a stronger positive residual in the subsurface. Again, the bias in SST asymmetry appears to be linked to the bias in the asymmetry of the subsurface temperature, as noted in Zhang et al. (2009).

To explore which phase of ENSO is the major source for the weaker residual in the SST and the subsurface in CMIP5 models, we investigate the spatial distribution of composite anomalies during two

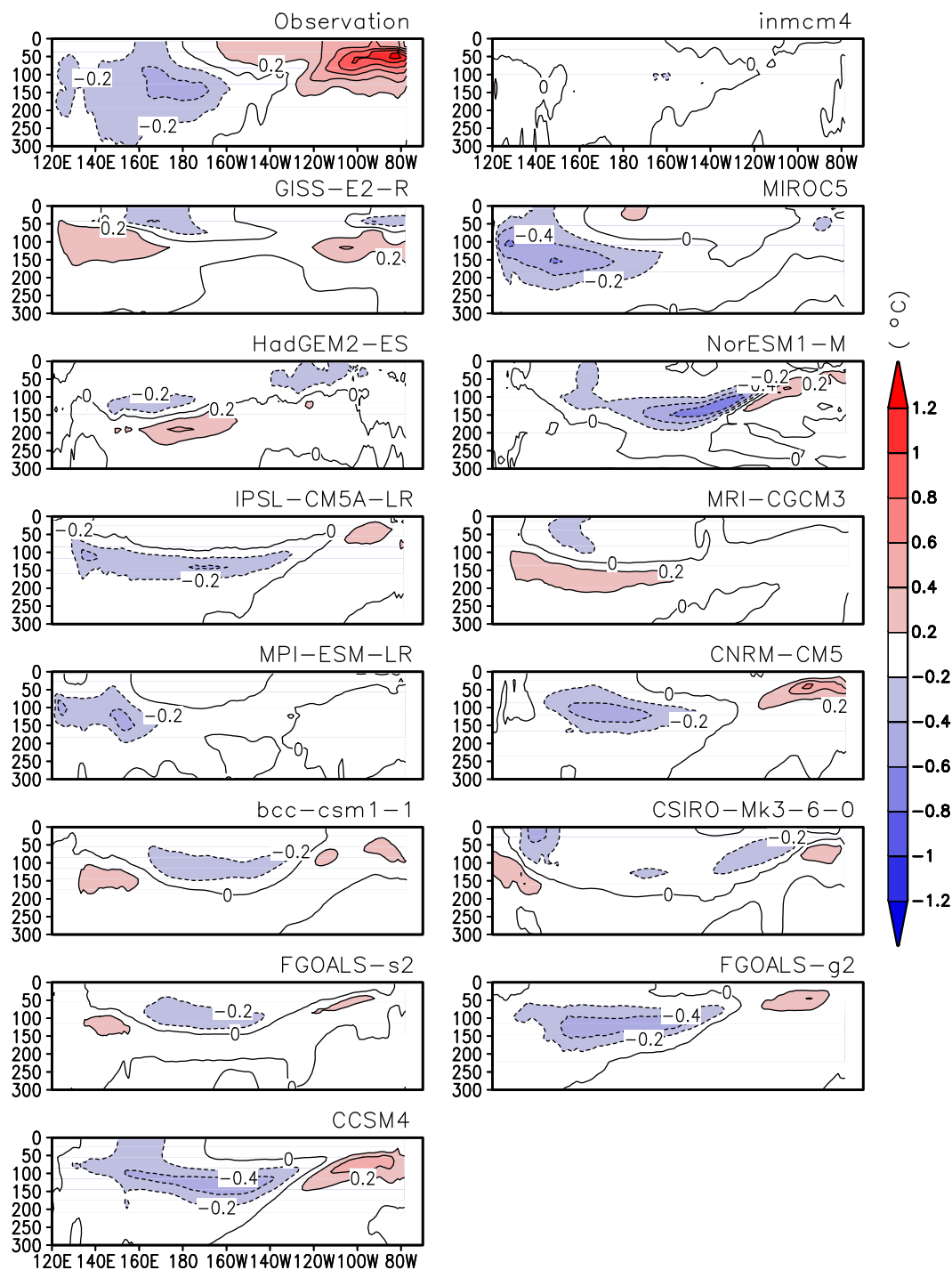


FIG. 3. The sum of the composite equatorial (5°S – 5°N) upper-ocean temperature anomalies between the two phases of ENSO from observations and CMIP5 coupled models. The length of data used in the calculation is 50 yr for all the models and SODA data (1950–99).

phases of ENSO. Figure 4 gives the spatial pattern of composite SST anomalies during the warm phase of ENSO. Observations show that the stronger positive SST anomalies associated with warm events are located over

the South American coast and the maximum value can reach about 1.6°C . Most models have a weaker SST warm anomaly over the eastern Pacific, and the underestimate of the warm SST anomaly is more serious

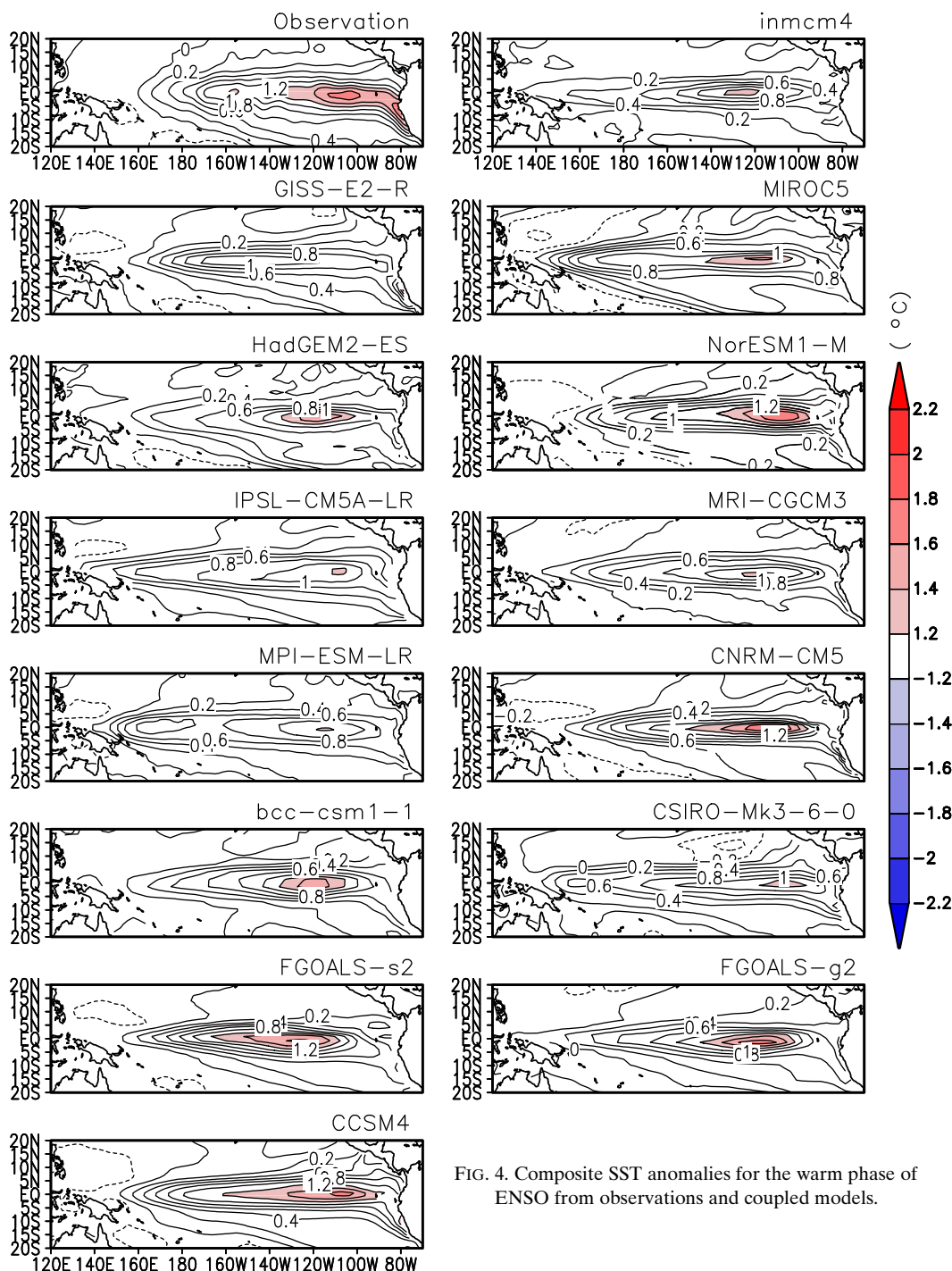


FIG. 4. Composite SST anomalies for the warm phase of ENSO from observations and coupled models.

in the coastal regions (100° – 80° W). The simulated maximum center is found to shift westward in many models. These biases contribute to the weak SST residual in the models (Fig. 2). The observed maximum center around 110° W is well captured in CCSM4, which has an enhanced warm anomaly over the coastal

regions that contributes to the increase in SST residual (Fig. 2).

The bias in the warm anomalies also shows up in the subsurface (Fig. 5). Consistent with the bias in the SST warm anomalies, most models have a weaker subsurface warm anomaly over the eastern Pacific and

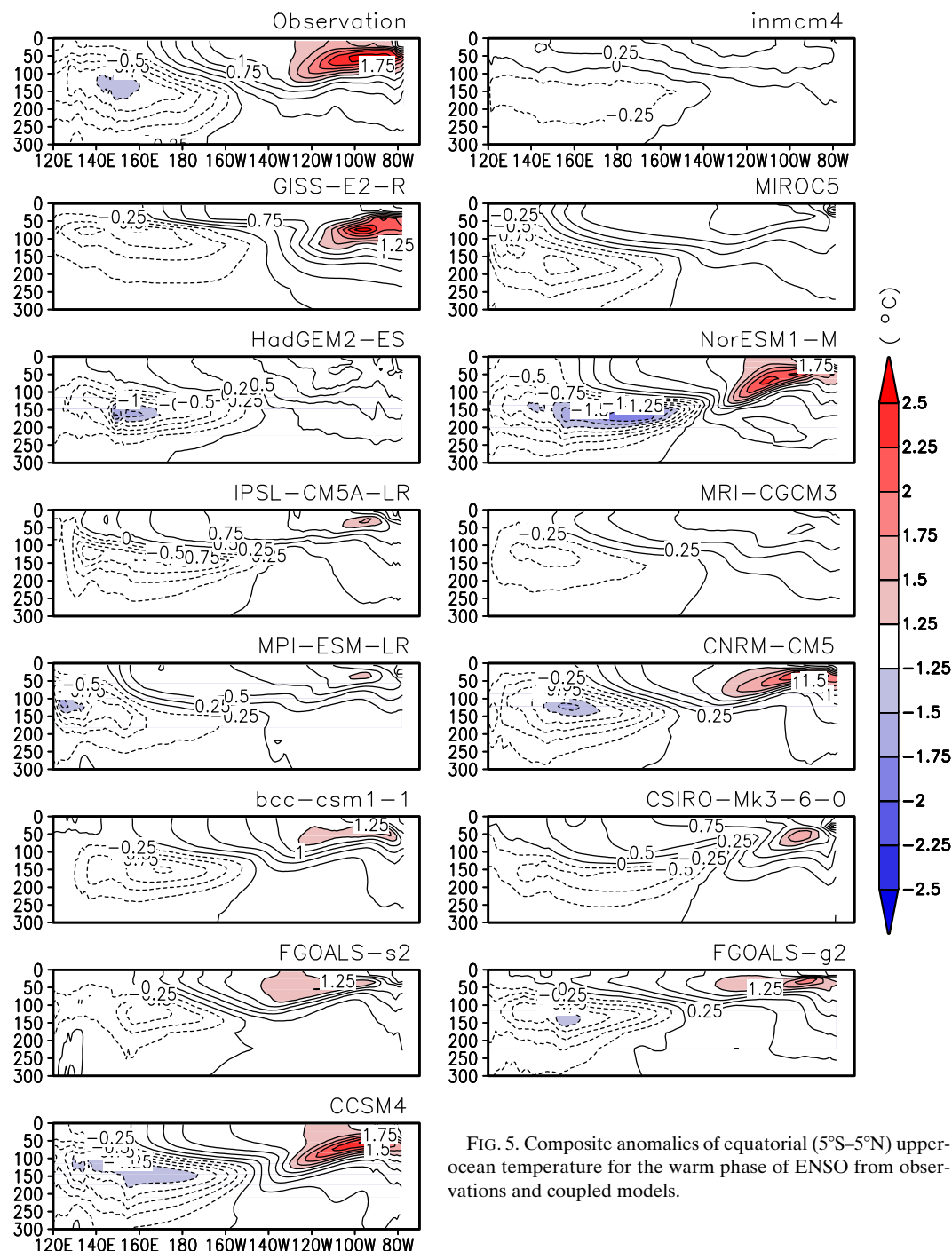


FIG. 5. Composite anomalies of equatorial (5°S – 5°N) upper-ocean temperature for the warm phase of ENSO from observations and coupled models.

the simulated maximum center is shifted westward. The better simulation of SST warm anomalies in CCSM4 is apparently associated with the improvement in the simulation of warm anomalies of subsurface temperature. Over the western Pacific, the underestimate of the negative anomalies in the subsurface is also evident in many models. The negative

anomalies in the subsurface over the western Pacific in NorESM1-M are much stronger and extend too far to the east during the warm phase, causing a stronger and eastward-extended negative asymmetry in this model (Fig. 3).

To better understand the cause for the underestimate of the ENSO asymmetry in CMIP5 coupled models, the

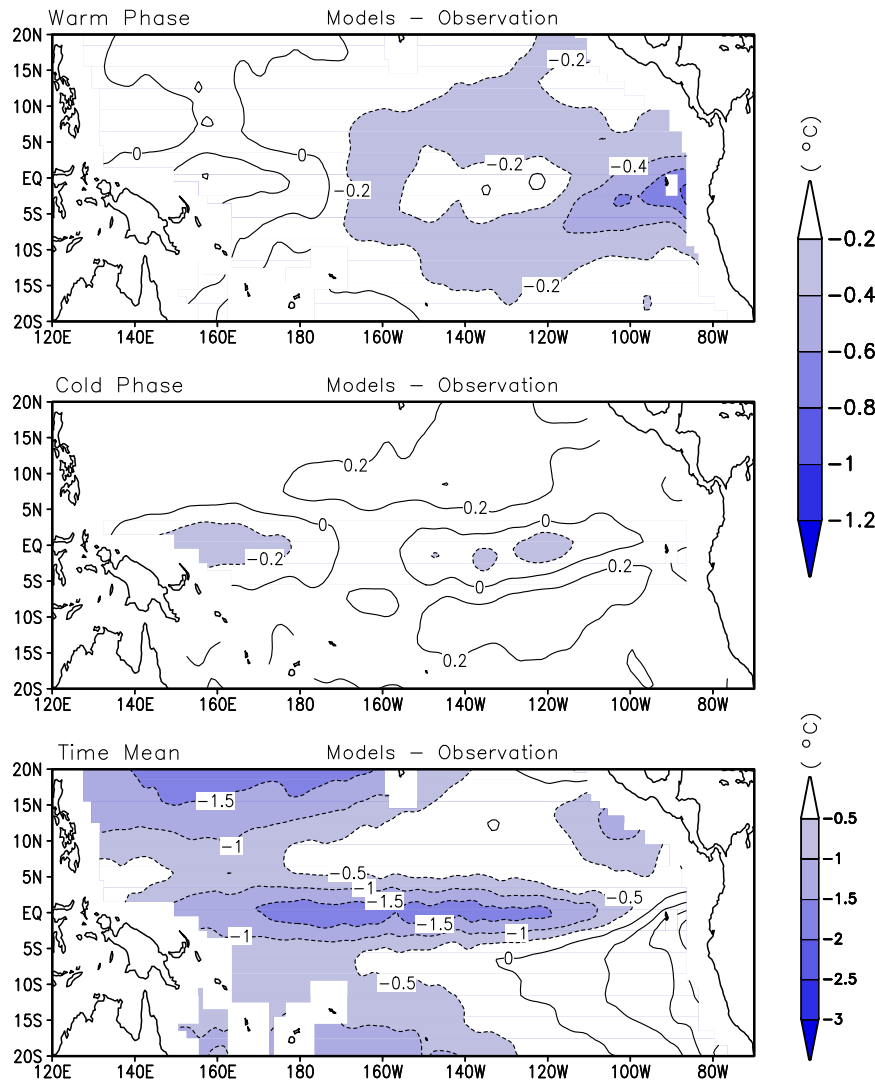


FIG. 6. (top) The difference between observations and ensemble mean composite SST anomalies for warm phase of ENSO, (middle) the difference between observations and ensemble mean composite SST anomalies for cold phase of ENSO, and (bottom) the difference between observations and ensemble mean SST annual climatology from 14 CMIP5 coupled models.

spatial map of the difference between models and observations for the composite SST anomalies during two phases of ENSO as well as time-mean SST is displayed in Fig. 6. Clearly, the underestimate of the warm anomalies is the major cause for the weaker ENSO asymmetry in CMIP5 coupled models, and the contribution from the bias during the cold phase of ENSO is small. We also note that CMIP5 models have a strong cold bias in mean SST state, a prevalent problem in coupled models (Sun et al. 2006; Zhang et al. 2009), which implies a possible link between the bias in mean SST state and the bias in ENSO asymmetry.

Figure 7 further shows the sum between the warm composite anomalies and cold composite anomalies in precipitation (shaded) and zonal wind stress (contours) from observations and coupled models. The observed precipitation is characterized by a strong positive asymmetry in the central and eastern Pacific and a strong negative asymmetry in the western Pacific, resulting from the westward shift during the cold phase compared to the warm phase (Zhang et al. 2009). The underestimate of the positive precipitation asymmetry over the central and eastern Pacific is prominent in the models. Consistent with the weak asymmetry in the

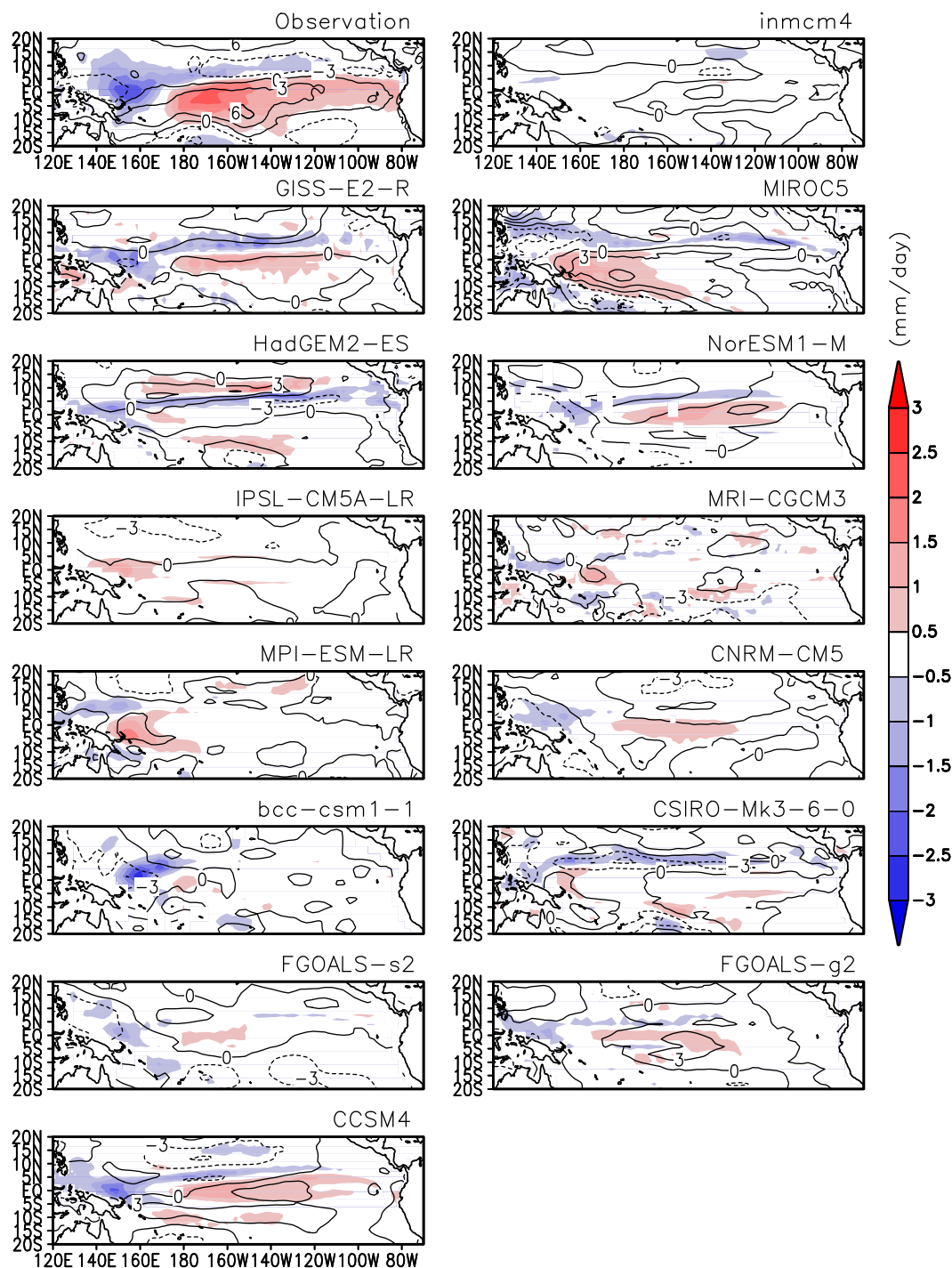


FIG. 7. The sum of the composite anomalies for the two phases of ENSO for precipitation (shaded) and zonal wind stress (contours) from observations and CMIP5 coupled models. The length of data used in the calculation is 50 yr for all the models, 30 yr for CMAP precipitation (1979–2008), and 50 yr for SODA zonal wind stress (1959–2008).

precipitation, the asymmetry in zonal wind stress is also weak in the coupled models, which is expected from the weak asymmetry in the subsurface temperature noted earlier.

b. Asymmetry in the AMIP runs

To understand whether the weaker asymmetry in precipitation and wind stress in CMIP5 coupled models

is a consequence of the corresponding SST fields or the cause of the latter, we perform the composite analysis from the corresponding AMIP runs of CMIP5 models that are forced by the observed SST boundary conditions. The AMIP runs involve subjecting the atmospheric component of CMIP5 coupled models to the observed ENSO SST variability and thus specifying the full ENSO asymmetry. The specification of the observed ENSO conditions in the AGCMs greatly increases the asymmetry in tropical Pacific rainfall, especially over the central Pacific, where the AMIP results are in much better agreement with observations than the results from coupled runs (Fig. 8). However, many models have a weaker precipitation asymmetry over the eastern Pacific. The NCAR model, which has proved to be the best model in simulating the ENSO asymmetry, is found to have a comparable precipitation asymmetry in the eastern Pacific. This suggests that the realistic simulation of precipitation asymmetry in the eastern Pacific may be an important factor for a better simulation of ENSO asymmetry.

Figure 9 shows a quantitative measure of the precipitation asymmetry over the eastern Pacific. The top panel shows results from the coupled runs and the bottom panel shows those from the corresponding AMIP runs. All the coupled models have a weaker precipitation asymmetry over the eastern Pacific. By comparison, the CCSM4 coupled model has the largest value of precipitation asymmetry. The increase in precipitation asymmetry from coupled runs to AMIP runs is also evident over the eastern Pacific. We also note that 9 of 14 AMIP models have a weaker precipitation asymmetry over the eastern Pacific even driven by the observed SST forcing. Two AMIP models (NorESM1-M and MRI-CGCM3) have a comparable precipitation asymmetry to the observed, and the other three AMIP models (GISS-E2-R, CCSM4, and BCC-CSM1-1) have a slightly larger precipitation asymmetry. The error of the weaker asymmetry in precipitation is apparently amplified in coupled runs as the coupled runs are found to have a much weaker precipitation asymmetry than their corresponding AMIP runs. There is a significant positive correlation (0.58) for the precipitation asymmetry averaged over the eastern Pacific between 14 AMIP runs and coupled runs. The weak precipitation asymmetry over the eastern Pacific is mainly due to the bias in the warm phase (Fig. 10). Out of 14 AMIP models, 9 have a weaker precipitation warm anomaly over the eastern Pacific. The precipitation warm anomaly is well captured in three models (HadGEM2-ES, MRI-CGCM3, and BCC-CSM1-1) and somewhat overestimated in the other two models (GISS-E2-R and CCSM4). Again, the corresponding coupled models have a much weaker

precipitation warm anomaly and all the coupled models underestimate the observed precipitation warm anomaly. This seems to indicate that the insufficient precipitation response to El Niño warming over the eastern Pacific is an intrinsic error of the majority of the atmospheric models. Further studies are needed to understand the cause of the bias in precipitation by exploring whether the model simply does not respond to the SST anomalies correctly in a local sense or there is a nonlocal influence from surface zonal stress, convergence, and the local reversal of the Walker circulation allowing or suppressing the ascent in the eastern Pacific.

Figure 11 further shows the spatial pattern during the warm phase for observations, the ensemble mean AMIP runs, and the differences between them. The left panel shows the precipitation and the right one the zonal wind stress. Consistent with the results shown in Fig. 10 (bottom), there is a weaker precipitation response over the eastern Pacific in the AMIP runs. Note that the precipitation warm anomalies in the AMIP runs are somewhat stronger over the central Pacific, and this positive bias is also reflected in the precipitation residual (Fig. 8), further confirming that the bias in the warm phase of ENSO is the major source for the bias in ENSO asymmetry. In contrast to observations, the precipitation response shows a less eastward extension (indicated by shaded values) in the AMIP runs during the warm phase. Linked to the precipitation response, the westerly wind (positive) anomaly is positioned too far to the west and shifts westward by about 10° (indicated by green lines shown in Fig. 11, top right and middle right). Similar to the precipitation difference, there is an obvious negative (weaker westerly wind) anomaly over the eastern Pacific and a positive (stronger westerly wind) anomaly over the central Pacific (Fig. 11, bottom right). The westward shift of the zonal wind stress warm anomalies in the AMIP runs may contribute to the weaker warm anomaly of subsurface temperature in most coupled models during the warm phase (Kang and Kug 2002). In addition to the westward shift of westerly wind anomaly, the significant easterly wind anomaly in the far eastern Pacific may also be responsible for the bias in subsurface temperature by inducing anomalous upwelling.

The CMIP5 AMIP runs are found to have biases in the mean zonal winds over the equatorial central and eastern Pacific and in the asymmetry in the central Pacific wind variability (Fig. 12). Of the 14 models, 11 have a stronger mean zonal wind in AMIP runs and the other 3 models (HadGEM2-ES, CCSM4, and CSIRO Mk3.6.0) have a mean wind comparable to the observed. Of the 14 models, 10 underestimate the observed positive skewness of central Pacific zonal winds in AMIP runs.

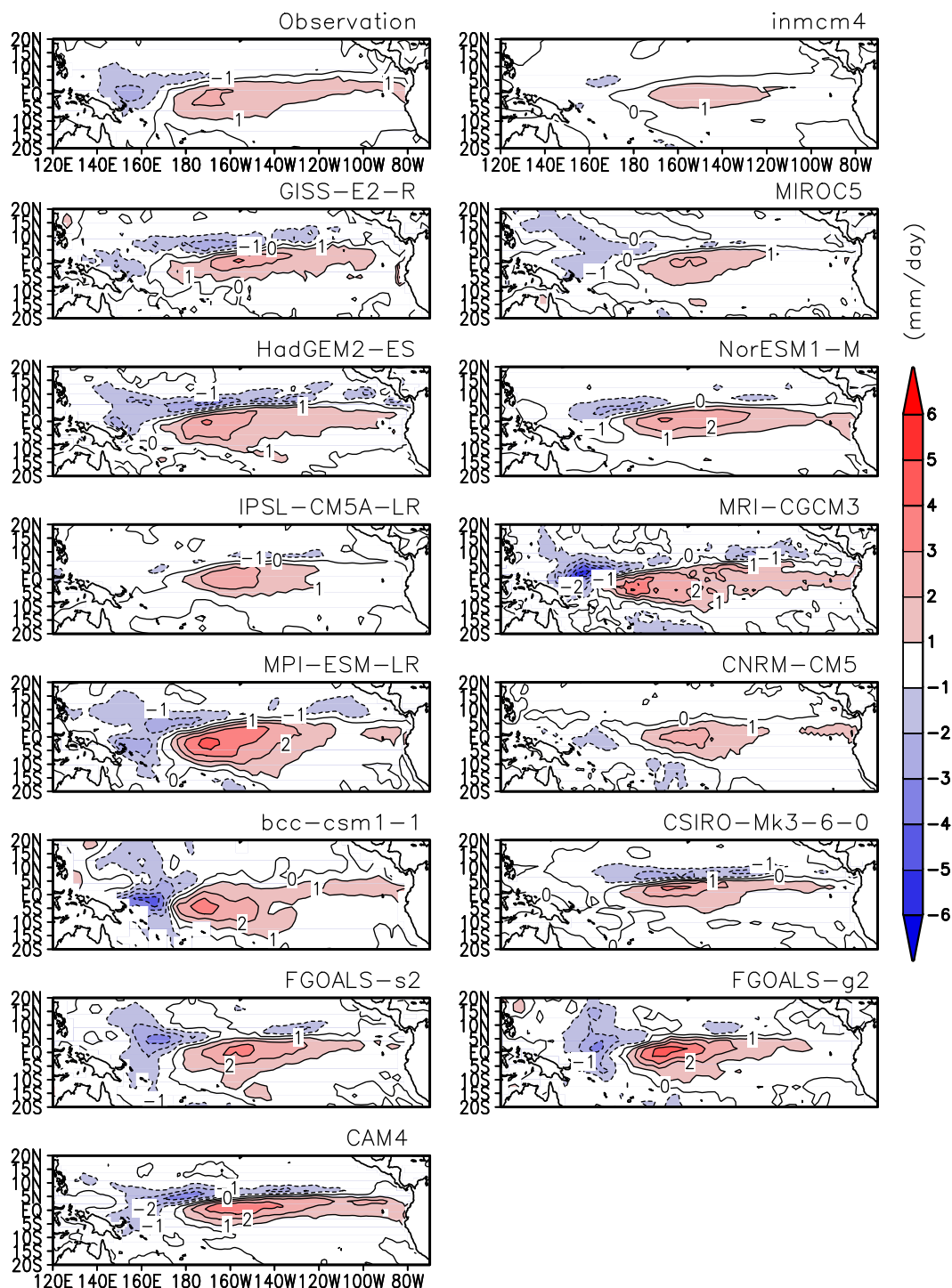


FIG. 8. The sum of the composite precipitation anomalies between the two phases of ENSO from observations and the corresponding AMIP runs of CMIP5 coupled models. The length of data used in the calculation is 30 yr for CMAP precipitation (1979–2008); 27 yr for the Community Atmosphere Model, version 4 (CAM4; 1979–2005); and 30 yr for the other models (1979–2008).

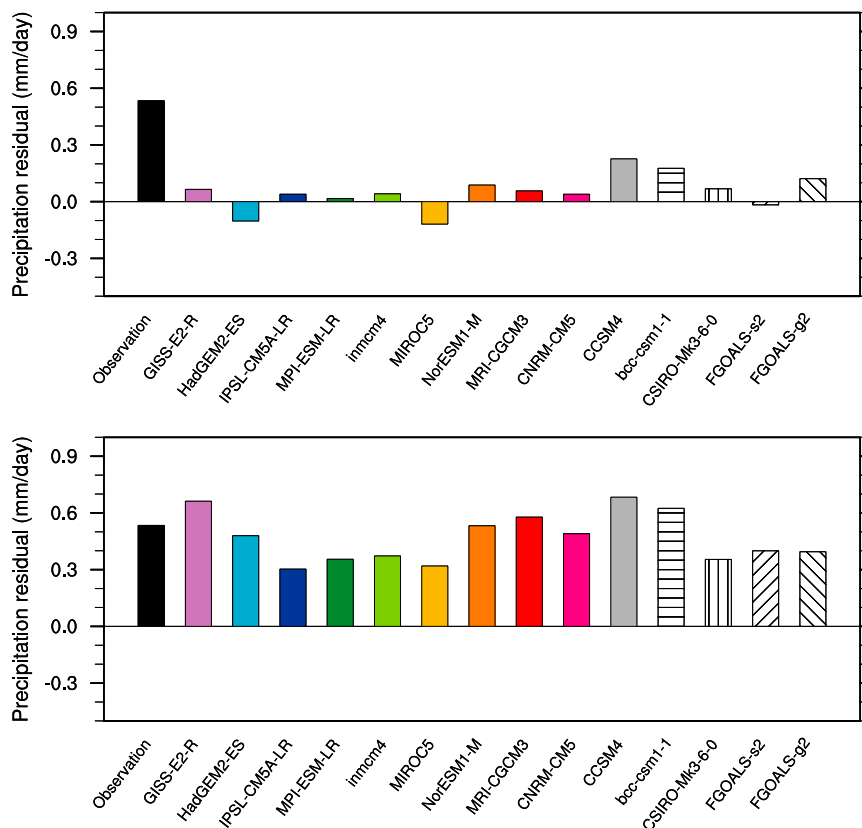


FIG. 9. The sum of the composite precipitation anomalies of the two phases of ENSO averaged over the eastern Pacific (120° – 70° W, 10° S– 10° N) from (top) CMIP5 coupled models and (bottom) the corresponding AMIP runs. The corresponding observational value is also included in the figures. The length of data used in the calculation is 30 yr for CMAP precipitation (1979–2008) and 50 yr for all the coupled models. The length of data used for AMIP runs is the same as in Fig. 8.

CNRM-CM5 and BCC-CSM1-1 have a better simulation of the observed wind skewness, while IPSL-CM5A-LR and MPI-ESM-LR have a stronger skewness in the zonal wind stress and the mean winds are also much stronger in these two models, especially in the latter. Generally, the ensemble mean results show that the AMIP runs have a stronger mean winds and a weaker skewness in the zonal winds.

The spatial map of time-mean zonal wind stress shows that there is a stronger mean wind in the models over most regions of the equatorial Pacific (Fig. 13). The bias in the mean wind (negative values) is more significant in the coastal regions (110° – 90° W, 0° – 10° N), where the mean precipitation is also much underestimated in the AMIP run. The stronger tropical winds are accompanied with excessive precipitation over much of the tropics, especially over the regions off the equator. More specifically, the mean precipitation difference is characterized with generally negative bias within the intertropical convergence zone (ITCZ) and with positive

bias elsewhere. The similar biases in winds and precipitation were also found in the previous CMIP3 AMIP runs (Lin 2007). There is a clear east–west asymmetry in the precipitation bias, and the resulting excessive zonal latent heating gradient associated with zonal precipitation gradient may drive the stronger winds in the model (Lin 2007). The results indicate that the climatological wind is an important cause of ENSO asymmetry. Specifically, the stronger mean winds will lead to a colder mean SST state that may suppress the increase of SST anomaly during the warm phase of ENSO but has less effect on the SST anomaly during the cold phase of ENSO. Probably associated with the dependence of the oceanic response on the mean SST state (McPhaden et al. 2011; Chung and Li 2013), this nonlinear effect of a colder mean state on the SST anomaly during the two phases of ENSO may be responsible for a weaker ENSO asymmetry. The biases in the surface winds from AMIP runs play a role in the ENSO asymmetry, which will be shown by the following numerical experiments.

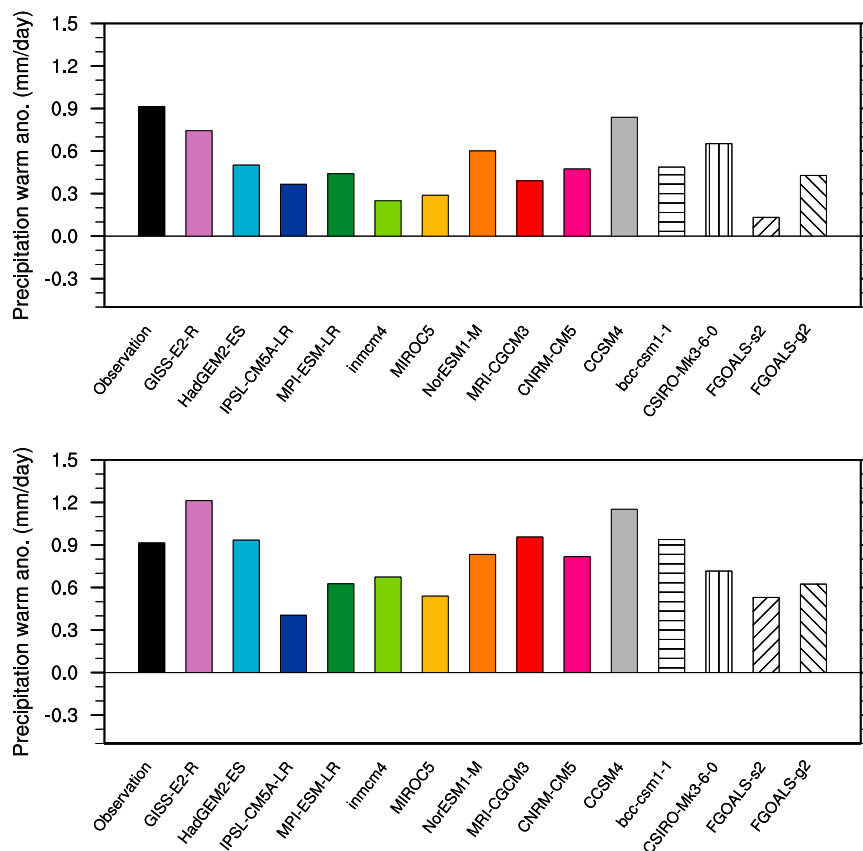


FIG. 10. Composite precipitation anomalies for the warm phase of ENSO averaged over the eastern Pacific (120° – 70° W, 10° S– 10° N) from (top) CMIP5 coupled models and (bottom) the corresponding AMIP runs.

c. Numerical experiments

To understand the biases of model winds associated with convection in AMIP runs on the ENSO asymmetry in CMIP5 coupled models, we use the NCAR Pacific basin model (Sun 2003; Sun et al. 2004; Sun and Zhang 2006) to perform numerical experiments. We conduct the forced ocean model experiments with the use of ensemble mean AMIP winds from 14 CMIP5 models and compare the results with those from the forced ocean runs driven by the observed wind stress. Four groups of numerical experiments combined with different climatology and interannual anomalies of winds in observations and ensemble mean AMIP runs of CMIP5 models are listed in Table 2. We first perform the forced ocean experiments with both climatology winds and interannual anomalies of winds from observations (experiment I). To understand the role of climatology winds in the models, we then replace the observed climatology winds by the modeled climatology winds but keep the observed interannual anomalies of winds unchanged in the forced experiments (experiment II).

Next, we use the actual AMIP model winds that include the simulated climatology and interannual anomalies to drive the ocean model, which will further explore the role of modeled interannual anomalies in the surface winds on ENSO asymmetry (experiment III). Last, to explore the role of observed climatology winds, we replace the modeled climatology winds with the observed climatology winds but keep the modeled interannual anomalies of winds to drive the ocean (experiment IV). These experiments are designed to probe the relative role of the bias in climatology winds and interannual variability of winds in AMIP runs in causing the underestimate of ENSO asymmetry in CMIP5 coupled runs.

Table 2 shows the standard deviation and skewness of the interannual variability in Niño-3 SST from four forced ocean runs. Driven by observed winds (experiment I), the model can well reproduce the observed skewness value of Niño-3 SST anomalies. The skewness value of 1.16 in experiment I is very close to the observed skewness value of 1.05 over the same 30-yr period. The results in the table show that the skewness from the run forced by full model winds (0.70 in

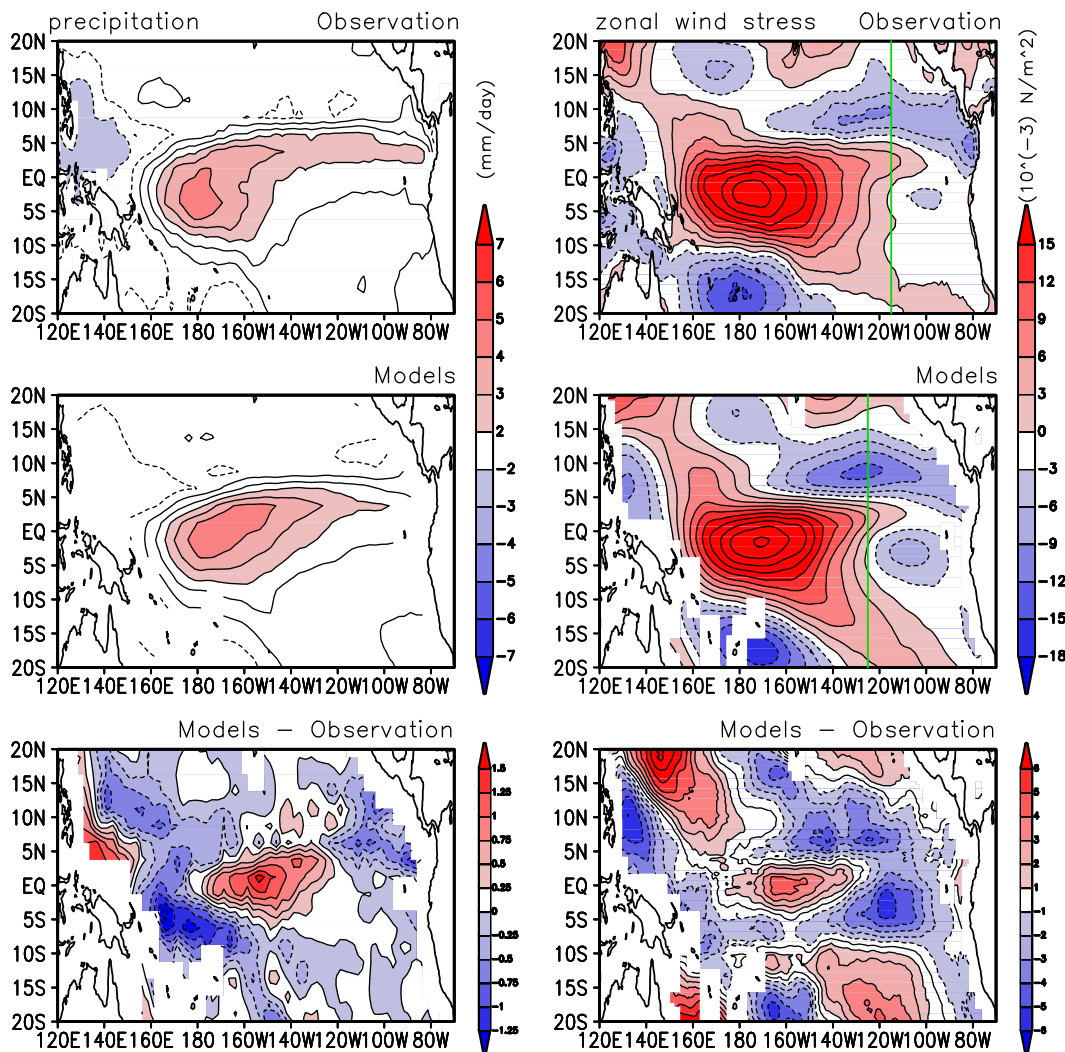


FIG. 11. (left) Warm phase precipitation anomalies and (right) zonal wind stress anomalies from observations, the ensemble mean of the model results, and their differences. Green lines indicate the positions that the equatorial westerly wind anomaly can reach. A total of 14 CMIP5 AMIP runs during the warm phase are used in calculating the ensemble mean. The length of observational data used in the calculation is 30 yr for CMAP precipitation and SODA zonal wind stress (1979–2008). The length of data used for AMIP runs is the same as in Fig. 8.

experiment III) is about 40% weaker than that from the run by the observed winds (1.16 in experiment I) accompanied by a weakened variability. By comparing the results from two cases that use the same observed wind anomaly but different wind climatology (experiments I and II), we find the bias in the modeled wind climatology is partially ($\sim 50\%$) responsible for the reduction in the ENSO asymmetry. The use of simulated wind interannual anomalies will further reduce the ENSO asymmetry, as the skewness in the run with full model winds is the smallest (experiment III). Interestingly, we note that the skewness in the case with observed wind climatology but keeping simulated wind interannual anomalies (experiment IV) is comparable to that in the run by the

observed full winds (experiment I), although the variability remains weak. The results from experiments III and IV indicate that the improvement in mean winds play a dominant role in improving the simulation of ENSO asymmetry.

The residual pattern of SST shows that there is a progressive decrease in the positive SST residual over the Niño-3 region from experiment I to experiment III (Fig. 14), consistent with the skewness value shown in Table 2. The decrease in the positive SST residual is more obvious in experiment III when full model winds are used. There is also a gradual westward shift in the positive SST residual, and the westward shift is also visible in the subsurface. The positive SST residual over the Niño-3

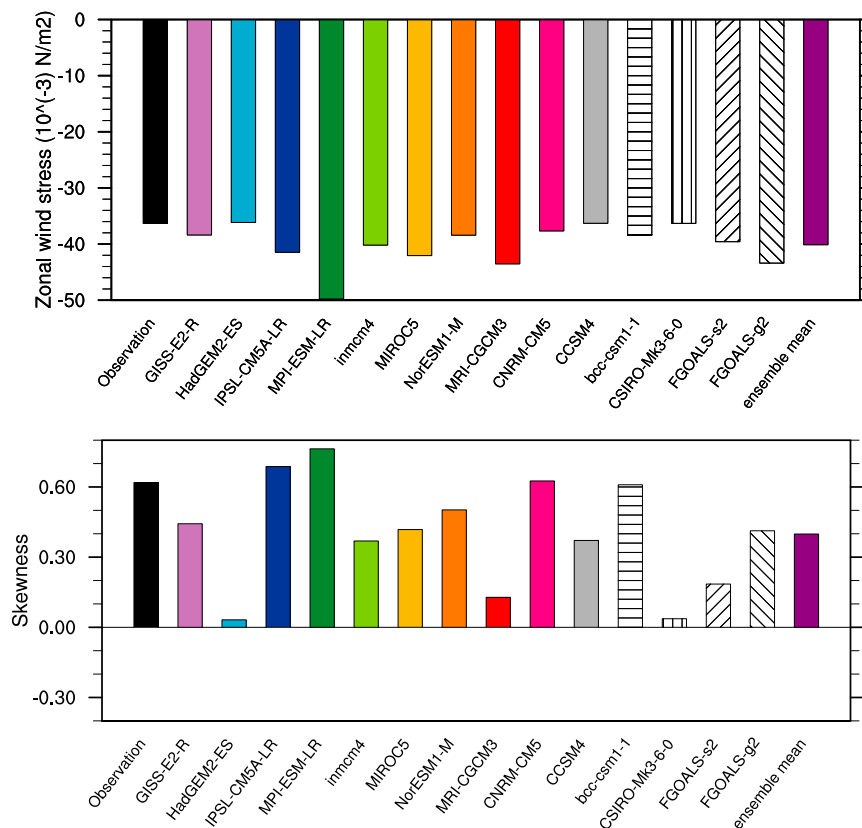


FIG. 12. (top) The time-mean zonal wind stress over the equatorial central and eastern Pacific (170°E–70°W, 5°S–5°N) and (bottom) the skewness of the interannual anomalies of the zonal wind stress over the central Pacific (160°E–140°W, 10°S–5°N) from observations and CMIP5 AMIP runs. The ensemble mean of the results from 14 AMIP runs is also included in the figure. Monthly anomalies are used to calculate the skewness. The length of observational data used in the calculation is 30 yr for SODA zonal wind stress (1979–2008). The length of data used for AMIP runs is the same as in Fig. 8.

region is greatly increased from experiment III to experiment IV when observed mean winds are used to replace the modeled mean winds, although there is a lack of evident positive SST residual over the coast regions (100°–80°W) in these two cases. Thus compared to observed wind anomalies, the wind anomalies in models can reduce the positive SST residual over the coast regions.

Figure 15 shows the spatial map of the composite anomalies of SST (left panel) and the equatorial upper-ocean temperature (right panel) during the warm phase of ENSO from four forced ocean experiments. The NCAR Pacific basin model used in this study reproduces the pattern of observed SST warm anomalies (Fig. 4). The simulated stronger SST warm anomalies in the run forced by observed winds (Fig. 15, top left) are located over the South American coast. The bias in the modeled wind climatology causes a slight westward shift of stronger SST warm anomaly but does not reduce the magnitude (Fig. 15, left, second row). Accompanied with a

weaker subsurface temperature warm anomaly (Fig. 15, right, third row), the westward shift of SST warm anomaly is more evident and the magnitude of SST warm anomaly becomes weaker if the bias in the interannual anomaly of modeled winds is also involved (Fig. 15, left, third row). The features of SST and subsurface temperature warm anomalies in the run forced by full model winds also exist in CMIP5 coupled models (Figs. 4, 5). The comparison between experiments III and IV shows that changing mean winds from models to observations alone can increase SST warm anomalies. Because of the use of the same model wind anomalies, the westward shift of SST warm anomaly is still evident in experiment IV. This is consistent with the lack of positive SST residual over the coast regions noted earlier (Fig. 14).

During the cold phase of ENSO (Fig. 16), bias in the modeled wind climatology somewhat increases the magnitude of cold SST anomalies over the Niño-3 region and thus reduces the SST skewness. The increase in

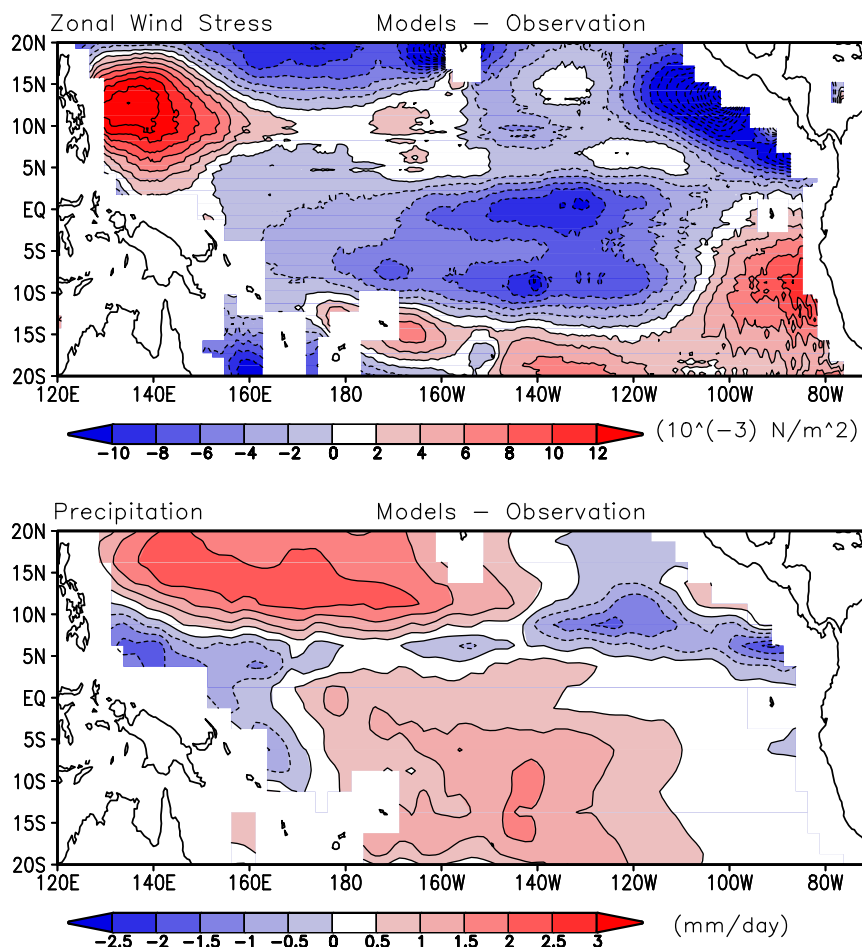


FIG. 13. (top) The difference between observations and the ensemble mean zonal wind stress annual climatology and (bottom) the difference between observations and ensemble mean precipitation annual climatology from 14 CMIP5 AMIP runs. The length of observational data used in the calculation is 30 yr for CMAP precipitation and SODA zonal wind stress (1979–2008). The length of data used for AMIP runs is the same as in Fig. 8.

cold SST anomaly magnitude is linked to the stronger cold subsurface temperature (Fig. 16, second row). Interestingly, the inclusion of wind anomalies from models is found to significantly reduce the magnitude of SST warm anomalies, but does not deteriorate the bias in cold SST anomalies (Fig. 16, third row). Instead, the cold

SST anomalies and subsurface temperature anomalies are comparable to those in the run forced by observed full winds. This also supports the previous analysis that the underestimate of the SST skewness in CMIP5 models is mostly due to bias in the warm phase. The experiment IV results show that the observed mean winds can reduce the

TABLE 2. Standard deviation and skewness of the interannual variability in Niño-3 SST from four forced ocean model experiments. The mean as well as the anomaly part of the surface winds used in these experiments are listed. The length of observed wind data used in the forced runs is 30 yr for SODA wind stress (1979–2008). The length of simulated wind data used is 27 yr for CAM4 (1979–2005) and 30 yr for other models (1979–2008).

Experiment (label in figures)	Surface wind stress		Statistics of Niño-3 SSTA	
	Climatology	Anomaly	Skewness	Standard deviation (°C)
Experiment I	Observation	Observation	1.16	0.75
Experiment II	CMIP5 AMIP ensemble	Observation	0.92	0.73
Experiment III	CMIP5 AMIP ensemble	CMIP5 AMIP ensemble	0.70	0.63
Experiment IV	Observation	CMIP5 AMIP ensemble	1.18	0.64

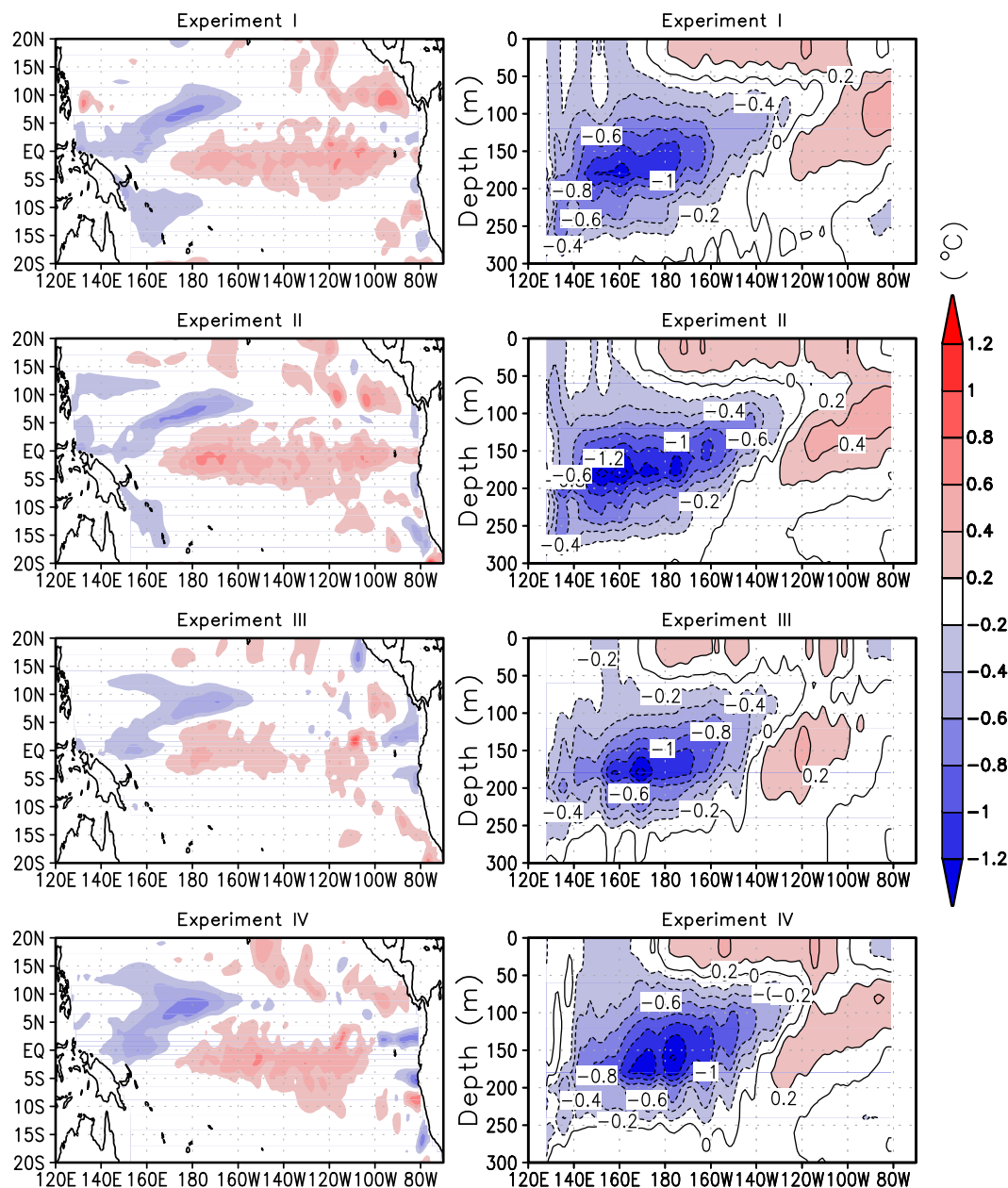


FIG. 14. The sum of the composite anomalies of the two phases of ENSO for (left) SST and (right) the equatorial (5°S–5°N) upper-ocean temperature in the four forced ocean experiments, as listed in Table 2.

cold SST anomalies, favoring an increase of SST skewness. Among the four runs, experiment IV has a more confined cold SST anomaly within the equatorial Pacific, while the other three runs show a more meridional extension of the cold SST anomalies, especially over the southern equatorial Pacific. The weakened cold SST anomalies over the Niño-3 region in experiment IV is linked to the reduction in cold subsurface temperature anomalies.

Figure 17 shows the time-mean SST difference and the equatorial upper-ocean temperature difference of

experiments II, III, and IV from experiment I. Compared to experiment I, there is a stronger cold SST over the cold-tongue regions in experiments II and III, in which the observed mean winds are replaced with mean winds from models. The subsurface temperature is also colder in these two cases that have a weaker SST skewness. By comparison, the SST and subsurface temperature in experiment IV are comparable to those in experiment I, since these cases use the same observed mean winds.

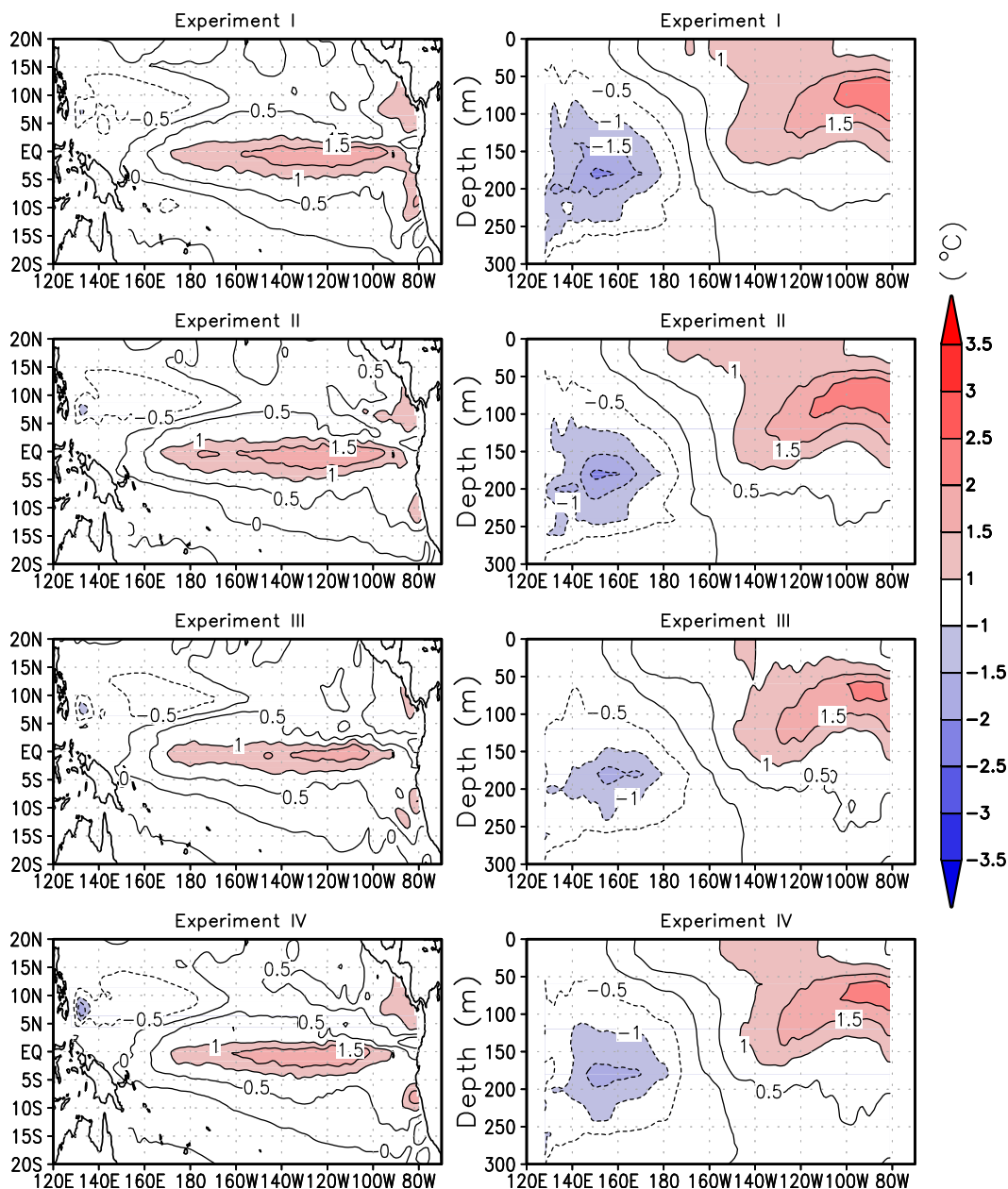


FIG. 15. Composite anomalies of (left) SST and (right) the equatorial (5°S – 5°N) upper-ocean temperature for the warm phase of ENSO in the four forced ocean experiments, as listed in Table 2.

Note that, different from experiment I, experiment IV uses the interannual anomalies of winds from models but still has a comparable SST skewness to the observed. This suggests that the mean SST state induced by mean winds is fundamentally important to the simulation of ENSO asymmetry and the bias in wind variability is secondary.

In general, the effect of the bias in interannual anomalies of modeled winds on ENSO asymmetry is mainly attributed to wind bias in the warm phase: a

westward shift of the zonal wind stress warm anomalies in the AMIP runs, linked to the insufficient precipitation response over the eastern Pacific during the warm phase (Fig. 11). These numerical experiments demonstrate that, when there is a colder mean SST state due to the stronger mean winds in models, the biases in interannual anomalies of winds from AMIP runs can weaken ENSO asymmetry by shifting SST warm anomalies westward and reducing their magnitude. When there is a warmer mean SST state or the model mean winds are the same as

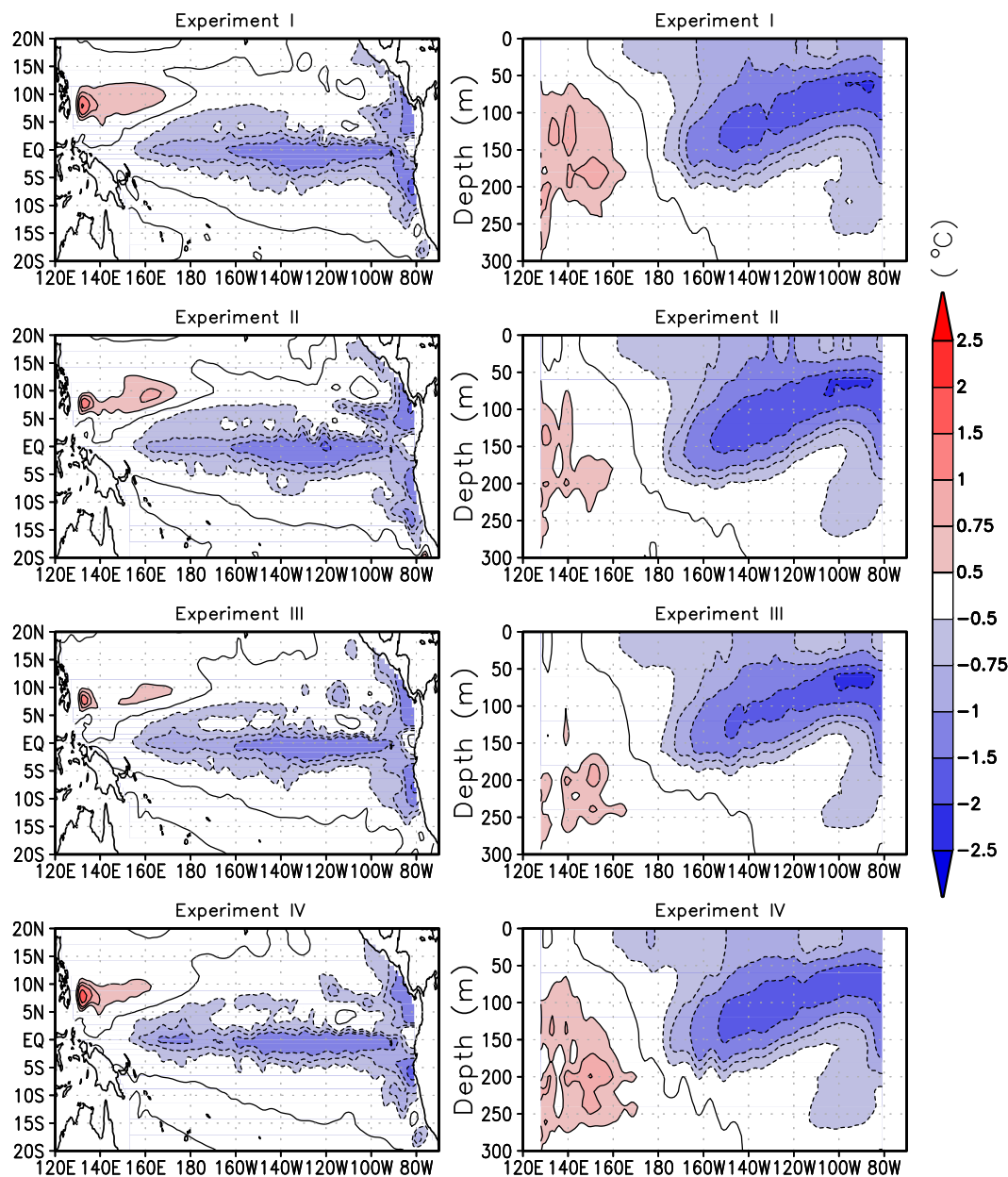


FIG. 16. As in Fig. 15, but for the cold phase of ENSO.

observations, the ENSO asymmetry can be as large as that in the run with observed full winds and the contribution to ENSO asymmetry from the bias in wind interannual variability is small.

4. Summary

In this study, we have evaluated the accuracy of CMIP5 coupled models in simulating the ENSO asymmetry and explored causes for bias in ENSO asymmetry in CMIP5 coupled models by analyzing the corresponding AMIP runs of CMIP5 coupled models and by conducting

forced ocean GCM experiments with the winds from CMIP5 AMIP runs.

Previous analysis of CMIP3 coupled models noted that, different from observations, most coupled models have a near-zero SST skewness in the tropical Pacific and a linear ENSO (van Oldenborgh et al. 2005; Sun et al. 2013). The present findings show that the underestimate of observed positive ENSO asymmetry measured by skewness is still a common problem in CMIP5 coupled models, although many models have comparable variance in Niño-3 SST with respect to

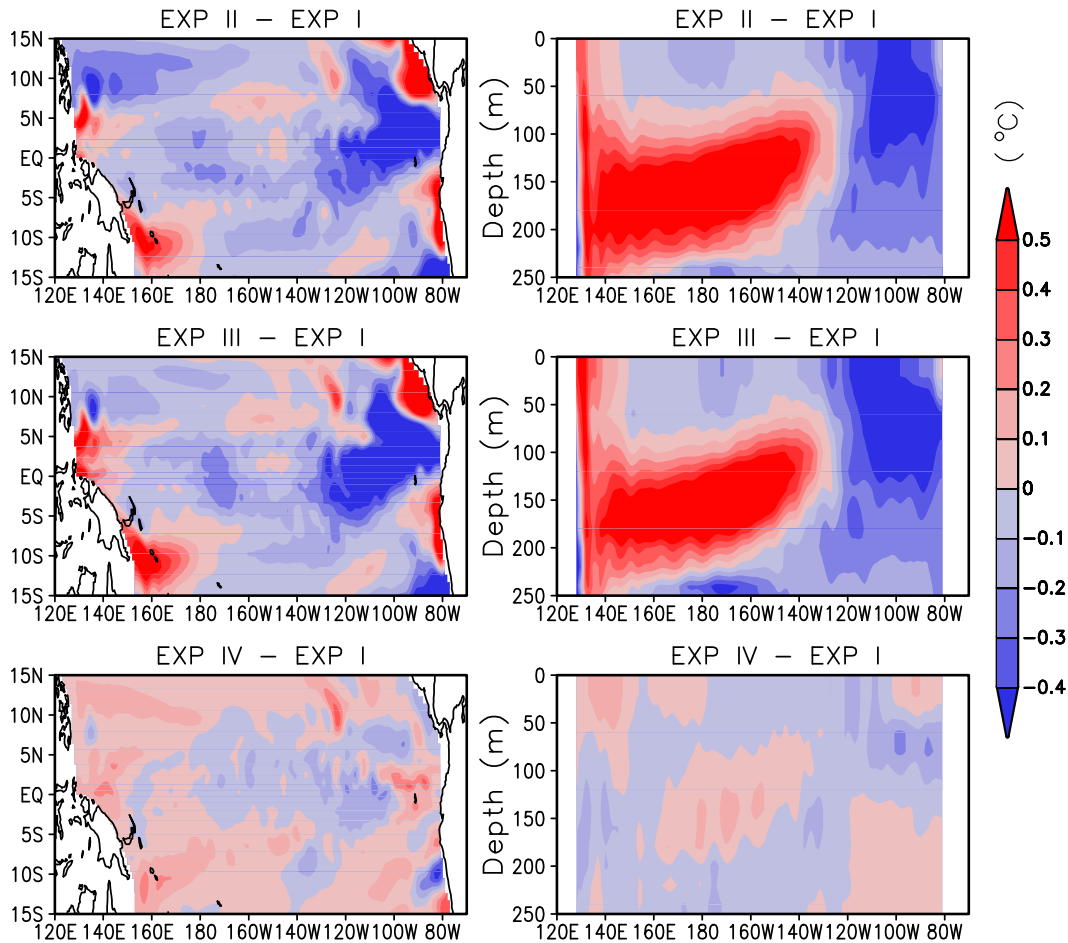


FIG. 17. (left) Time-mean SST difference and (right) the equatorial (5°S – 5°N) upper-ocean temperature difference of experiments II, III, and IV from experiment I.

observations, a significant improvement over CMIP3. When the asymmetry is measured by the SST residual between the two phases of ENSO, all the models are also found to have a weaker ENSO asymmetry than observations. It is notable that CMIP5 coupled models have a significant cold bias in the mean SST, as seen in many coupled models (Sun et al. 2006; Zhang et al. 2009). The weak ENSO asymmetry in CMIP5 models has corresponding signatures in biases in zonal wind stress, precipitation, and subsurface temperatures, which are also too symmetrical with respect to ENSO phases. The composite analysis indicates that the weaker asymmetry of ENSO in CMIP5 coupled models is largely a consequence of the bias from El Niño events. The SST warm anomalies over the far eastern Pacific are found to be weaker in the coupled models than in observations and the simulated maximum warm SST center over the eastern Pacific shifts westward. Most models also have a weaker subsurface temperature warm anomaly over the eastern Pacific and the maximum center shifts westward.

The asymmetry in the precipitation and zonal wind stress from the corresponding AMIP runs are first analyzed to understand the causes for the weaker ENSO asymmetry (or the weaker El Niño events) in CMIP5 coupled models. We found that, mainly because of the weaker precipitation response to El Niño warming, most models have a weaker precipitation asymmetry over the eastern Pacific even driven by the observed SST forcing. This bias is further amplified in the coupled models that have a much weaker precipitation asymmetry over the eastern Pacific. During the warm phase, the weaker precipitation response over the eastern Pacific is accompanied by a stronger precipitation response over the central Pacific and linked to a westward shift of convection in the AMIP runs along with a clear westward shift of westerly wind anomaly. A westward shift of zonal wind stress during the warm phase in the AMIP runs may play a role in the weaker subsurface temperature warm anomalies in the coupled models (Kang and Kug 2002). Using two different coupled models to

examine the sensitivity of ENSO amplitude to the convection scheme parameters, Watanabe et al. (2011) and Kim et al. (2011) showed that the parameter change in the cumulus parameterization shifts the position of the precipitation anomalies and the zonal wind stress also shifts accordingly. The increased eastern Pacific precipitation tends to shift the wind stress anomalies to the east. Closer to the eastern Pacific, the wind stress forcing more effectively deepens the thermocline over the eastern Pacific. Watanabe et al. (2011) have also showed that the subsurface temperature anomalies over the eastern Pacific are much stronger when the zonal wind stress shifts to the east. This is consistent with what we see from CCSM4. The NCAR model, identified as the best model in simulating ENSO asymmetry, has a realistic simulation of subsurface temperature warm anomalies associated with sufficient precipitation response over the eastern Pacific in the AMIP run. An enhanced precipitation response over the eastern Pacific during the warm phase is essential to the improvement in the simulation of ENSO asymmetry in CMIP5 models, consistent with the previous findings of Zhang et al. (2009).

We also find that most AMIP models have a stronger time-mean zonal wind over the equatorial central and eastern Pacific and underestimate the observed positive skewness of zonal winds in the central Pacific. The bias in the mean zonal winds is more prominent in the coastal regions over the eastern Pacific and the southern equatorial Pacific, where the bias in mean precipitation is also evident in the AMIP runs. The mean precipitation bias shows an east–west asymmetry. The latent heating asymmetry associated with the stronger zonal precipitation gradient may generate the stronger zonal pressure gradient force, which then enhances the trade winds in the model (Lin 2007).

To understand the effect of the bias in the mean and interannual variability of winds on ENSO asymmetry, forced ocean model experiments with the use of AMIP winds are performed. These results are compared to those from the experiments forced by observed winds. The numerical experiments show that, when there is a colder mean SST state because of the stronger mean winds in models, the biases in interannual anomalies of winds from AMIP runs can weaken ENSO asymmetry by shifting SST warm anomalies westward and reducing the magnitude. This is consistent with what we have seen in CMIP5 coupled models. The results from the run with full model winds confirm that the bias in the SST anomalies during the warm phase is found to be the major cause for the reduction in ENSO asymmetry. We note that, with a warmer mean SST state or when the mean winds in models are the same as observations, the contribution to ENSO asymmetry from wind interannual

variability bias is negligible. Also, ENSO asymmetry is increased mainly because of the increase of SST warm anomalies. The results are consistent with those from an analytical model that the amplitude of warm events increases with enhanced radiative heating (Liang et al. 2012). This may also be useful to explain why coupled models tend to have a weaker ENSO asymmetry, given that the excessive cold tongue is still the problem in coupled models (Sun et al. 2006). These findings highlight the importance of a warmer mean SST state for ENSO asymmetry. Further studies are needed to explore this possible link.

To the extent a colder mean state of the ocean causes a weaker ENSO asymmetry and to the extent this colder mean state is mainly a consequence of the stronger zonal wind from the AMIP runs, our analysis pinpoints the causes of the weaker ENSO asymmetry in the coupled models to the stronger time-mean winds over the tropical Pacific in the stand-alone atmosphere model. Note that we have fully considered the momentum forcing (both zonal and meridional components) from AMIP model winds in the experimental design as an attempt to reveal the role of the bias in model winds more realistically. We have also performed additional ocean model experiments in which only zonal wind stress biases are considered. The results are found to be similar, suggesting a minor role of the biases in meridional wind stress, as also noted in previous studies (McCreary 1976; Zhang and McPhaden 2006; Zhu et al. 2007).

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REFERENCES

- An, S.-I., 2009: A review of interdecadal changes in the nonlinearity of the El Niño–Southern Oscillation. *Theor. Appl. Climatol.*, **97**, 29–40, doi:10.1007/s00704-008-0071-z.
- , and F.-F. Jin, 2004: Nonlinearity and asymmetry of ENSO. *J. Climate*, **17**, 2399–2412, doi:10.1175/1520-0442(2004)017<2399:NAAOE>2.0.CO;2.
- , Y.-G. Ham, J.-S. Kug, F.-F. Jin, and I.-S. Kang, 2005: El Niño–La Niña asymmetry in the Coupled Model Intercomparison Project simulations. *J. Climate*, **18**, 2617–2627, doi:10.1175/JCLI3433.1.
- Burgers, G., and D. B. Stephenson, 1999: The “normality” of El Niño. *Geophys. Res. Lett.*, **26**, 1027–1030, doi:10.1029/1999GL900161.
- Carton, J. A., and B. S. Giese, 2008: A reanalysis of ocean climate using Simple Ocean Data Assimilation (SODA). *Mon. Wea. Rev.*, **136**, 2999–3017, doi:10.1175/2007MWR1978.1.

- , G. Chepurin, X. Cao, and B. Giese, 2000: A simple ocean data assimilation analysis of the global upper ocean 1950–95. Part I: Methodology. *J. Phys. Oceanogr.*, **30**, 294–309, doi:[10.1175/1520-0485\(2000\)030<0294:ASODAA>2.0.CO;2](#).
- Chung, P.-H., and T. Li, 2013: Interdecadal relationship between the mean state and El Niño types. *J. Climate*, **26**, 361–379, doi:[10.1175/JCLI-D-12-00106.1](#).
- Deser, C., and Coauthors, 2012: ENSO and Pacific decadal variability in the Community Climate System Model version 4. *J. Climate*, **25**, 2622–2651, doi:[10.1175/JCLI-D-11-00301.1](#).
- Gent, P. R., and M. A. Cane, 1989: A reduced gravity, primitive equation model of the upper equatorial ocean. *J. Comput. Phys.*, **81**, 444–480, doi:[10.1016/0021-9991\(89\)90216-7](#).
- , and Coauthors, 2011: The Community Climate System Model version 4. *J. Climate*, **24**, 4973–4991, doi:[10.1175/2011JCLI4083.1](#).
- Hannachi, A., D. Stephenson, and K. Sperber, 2003: Probability-based methods for quantifying nonlinearity in the ENSO. *Climate Dyn.*, **20**, 241–256, doi:[10.1007/s00382-002-0263-7](#).
- Hoerling, M. P., A. Kumar, and M. Zhong, 1997: El Niño, La Niña, and the nonlinearity of their teleconnections. *J. Climate*, **10**, 1769–1786, doi:[10.1175/1520-0442\(1997\)010<1769:ENOLNA>2.0.CO;2](#).
- Jin, F.-F., S.-I. An, A. Timmermann, and J. Zhao, 2003: Strong El Niño events and nonlinear dynamical heating. *Geophys. Res. Lett.*, **30**, 1120, doi:[10.1029/2002GL016356](#).
- Kang, I.-S., and J.-S. Kug, 2002: El Niño and La Niña sea surface temperature anomalies: Asymmetry characteristics associated with their wind stress anomalies. *J. Geophys. Res.*, **107**, 4372, doi:[10.1029/2001JD000393](#).
- Kiladis, G. N., and H. Diaz, 1989: Global climatic anomalies associated with extremes in the Southern Oscillation. *J. Climate*, **2**, 1069–1090, doi:[10.1175/1520-0442\(1989\)002<1069:GCAAW>2.0.CO;2](#).
- Kim, D., Y.-S. Jang, D.-H. Kim, Y.-H. Kim, M. Watanabe, F.-F. Jin, and J.-S. Kug, 2011: El Niño–Southern Oscillation sensitivity to cumulus entrainment in a coupled general circulation model. *J. Geophys. Res.*, **116**, D22112, doi:[10.1029/2011JD016526](#).
- Larkin, N. K., and D. E. Harrison, 2005: On the definition of El Niño and associated seasonal average U.S. weather anomalies. *Geophys. Res. Lett.*, **32**, L13705, doi:[10.1029/2005GL022738](#).
- Liang, J., X.-Q. Yang, and D.-Z. Sun, 2012: The effect of ENSO events on the tropical Pacific mean climate: Insights from an analytical model. *J. Climate*, **25**, 7590–7606, doi:[10.1175/JCLI-D-11-00490.1](#).
- Lin, J.-L., 2007: The double-ITCZ problem in IPCC AR4 coupled GCMs: Ocean–atmosphere feedback analysis. *J. Climate*, **20**, 4497–4525, doi:[10.1175/JCLI4272.1](#).
- McCreary, J., 1976: Eastern tropical ocean response to changing wind systems: With application to El Niño. *J. Phys. Oceanogr.*, **6**, 632–645, doi:[10.1175/1520-0485\(1976\)006<0632:ETORTC>2.0.CO;2](#).
- McPhaden, M. J., T. Lee, and D. McClurg, 2011: El Niño and its relationship to changing background conditions in the tropical Pacific Ocean. *Geophys. Res. Lett.*, **38**, L15709, doi:[10.1029/2011GL048275](#).
- Neale, R. B., J. H. Richter, and M. Jochum, 2008: The impact of convection on ENSO: From a delayed oscillator to a series of events. *J. Climate*, **21**, 5904–5924, doi:[10.1175/2008JCLI2244.1](#).
- Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**, 4407, doi:[10.1029/2002JD002670](#).
- Rodgers, K. B., P. Friederichs, and M. Latif, 2004: Tropical Pacific decadal variability and its relation to decadal modulation of ENSO. *J. Climate*, **17**, 3761–3774, doi:[10.1175/1520-0442\(2004\)017<3761:TPDVAI>2.0.CO;2](#).
- Ropelewski, C. F., and M. S. Halpert, 1987: Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation. *Mon. Wea. Rev.*, **115**, 1606–1626, doi:[10.1175/1520-0493\(1987\)115<1606:GARSPP>2.0.CO;2](#).
- Schopf, P. S., and R. J. Burgman, 2006: A simple mechanism for ENSO residuals and asymmetry. *J. Climate*, **19**, 3167–3179, doi:[10.1175/JCLI3765.1](#).
- Su, J., R. Zhang, T. Li, X. Rong, J.-S. Kug, and C.-C. Hong, 2010: Causes of the El Niño and La Niña amplitude asymmetry in the equatorial eastern Pacific. *J. Climate*, **23**, 605–617, doi:[10.1175/2009JCLI2894.1](#).
- Sun, D.-Z., 1997: El Niño: A coupled response to radiative heating? *Geophys. Res. Lett.*, **24**, 2031–2034, doi:[10.1029/97GL01960](#).
- , 2003: A possible effect of an increase in the warm-pool SST on the magnitude of El Niño warming. *J. Climate*, **16**, 185–205, doi:[10.1175/1520-0442\(2003\)016<0185:APEOAI>2.0.CO;2](#).
- , and T. Zhang, 2006: A regulatory effect of ENSO on the time-mean thermal stratification of the equatorial upper ocean. *Geophys. Res. Lett.*, **33**, L07710, doi:[10.1029/2005GL025384](#).
- , and F. Bryan, Eds., 2010: *Climate Dynamics: Why Does Climate Vary?* *Geophys. Monogr.*, Vol. 189, Amer. Geophys. Union, 216 pp.
- , T. Zhang, and S.-I. Shin, 2004: The effect of subtropical cooling on the amplitude of ENSO: A numerical study. *J. Climate*, **17**, 3786–3798, doi:[10.1175/1520-0442\(2004\)017<3786:TEOSCO>2.0.CO;2](#).
- , and Coauthors, 2006: Radiative and dynamical feedbacks over the equatorial cold tongue: Results from nine atmospheric GCMs. *J. Climate*, **19**, 4059–4074, doi:[10.1175/JCLI3835.1](#).
- , T. Zhang, Y. Sun, and Y. Yu, 2014: Rectification of El Niño–Southern Oscillation into climate anomalies of decadal and longer time scales: Results from forced ocean GCM experiments. *J. Climate*, **27**, 2545–2561, doi:[10.1175/JCLI-D-13-00390.1](#).
- Sun, F., and J.-Y. Yu, 2009: A 10–15-year modulation cycle of ENSO intensity. *J. Climate*, **22**, 1718–1735, doi:[10.1175/2008JCLI2285.1](#).
- Sun, Y., D.-Z. Sun, L. X. Wu, and F. Wang, 2013: Western Pacific warm pool and ENSO asymmetry in CMIP3 models. *Adv. Atmos. Sci.*, **30**, 940–953, doi:[10.1007/s00376-012-2161-1](#).
- Taylor, K. E., R. J. Stouffer, and G. A. Meehl, 2012: An overview of CMIP5 and the experiment design. *Bull. Amer. Meteor. Soc.*, **93**, 485–498, doi:[10.1175/BAMS-D-11-00094.1](#).
- van Oldenborgh, G. J., S. Philip, and M. Collins, 2005: El Niño in a changing climate: A multi-model study. *Ocean Sci.*, **1**, 81–95, doi:[10.5194/os-1-81-2005](#).
- Vialard, J., C. Menkes, J.-P. Boulanger, P. Delecluse, E. Guilyardi, M. J. McPhaden, and G. Madec, 2001: A model study of oceanic mechanisms affecting equatorial Pacific sea surface temperature during the 1997–98 El Niño. *J. Phys. Oceanogr.*, **31**, 1649–1675, doi:[10.1175/1520-0485\(2001\)031<1649:AMSOOM>2.0.CO;2](#).
- Watanabe, M., M. Chikira, Y. Imada, and M. Kimoto, 2011: Convective control of ENSO simulated in MIROC. *J. Climate*, **24**, 543–562, doi:[10.1175/2010JCLI3878.1](#).
- Xie, P., and P. A. Arkin, 1997: Global precipitation: A 17-year monthly analysis based on gauge observations, satellite estimates,

- and numerical model outputs. *Bull. Amer. Meteor. Soc.*, **78**, 2539–2558, doi:[10.1175/1520-0477\(1997\)078<2539:GPAYMA>2.0.CO;2](https://doi.org/10.1175/1520-0477(1997)078<2539:GPAYMA>2.0.CO;2).
- Zhang, G. J., and N. A. McFarlane, 1995: Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian Climate Centre general circulation model. *Atmos.–Ocean*, **33**, 407–446, doi:[10.1080/07055900.1995.9649539](https://doi.org/10.1080/07055900.1995.9649539).
- Zhang, T., D.-Z. Sun, R. Neale, and P. J. Rasch, 2009: An evaluation of ENSO asymmetry in the Community Climate System Models: A view from the subsurface. *J. Climate*, **22**, 5933–5961, doi:[10.1175/2009JCLI2933.1](https://doi.org/10.1175/2009JCLI2933.1).
- , M. P. Hoerling, J. Perlwitz, D.-Z. Sun, and D. Murray, 2011: Physics of U.S. surface temperature response to ENSO. *J. Climate*, **24**, 4874–4887, doi:[10.1175/2011JCLI3944.1](https://doi.org/10.1175/2011JCLI3944.1).
- , J. Perlwitz, and M. P. Hoerling, 2014: What is responsible for the strong observed asymmetry in teleconnections between El Niño and La Niña? *Geophys. Res. Lett.*, **41**, 1019–1025, doi:[10.1002/2013GL058964](https://doi.org/10.1002/2013GL058964).
- Zhang, X., and M. J. McPhaden, 2006: Wind stress variations and interannual sea surface temperature anomalies in the eastern equatorial Pacific. *J. Climate*, **19**, 226–241, doi:[10.1175/JCLI3618.1](https://doi.org/10.1175/JCLI3618.1).
- Zhu, J., Z. Sun, and G. Zhou, 2007: A note on the role of meridional wind stress anomalies and heat flux in ENSO simulations. *Adv. Atmos. Sci.*, **24**, 729–738, doi:[10.1007/s00376-007-0729-y](https://doi.org/10.1007/s00376-007-0729-y).